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IGR TRANSLITERATION OF RUSSIAN

The AGI Translation Office has adopted the essential features of Cyrillic transliteration recommended by the U. S. Department of the Interior, Board on Geographic Names, Washington D. C.

However, the AGI Translation Office recommends the following modifications:

1. Ye initially, after vowels, and after Ъ, Ь Customary usage calls for "ie" in many names, e. g., SOVIET KIEV, DNEPER, etc.; or "ye", e. g., BYELORUSSIA, where "e" follows consonants. "e" with dieresis in Russian should be given as "yo".
2. Omitted if preceding a "y", for example, Arkhangelsky (not "iy"; not "ii").
3. Generally omitted.

NOTE: Well-known place and personal names that have wide acceptance will be used. Some translations may include elements of previous German transliteration from the Russian; this occurs in IGR most commonly in maps and lists of references. The reader's attention is called to the following variations between German and English systems which may cause confusion when trying to check back to original Russian sources.

Alphabet	transliteration	
А	а	a
Б	б	b
В	в	v
Г	г	g
Д	д	d
Е	е	e, ye ⁽¹⁾
Ё	ё	ë, ye
Ж	ж	zh
З	з	z
И	и	i ⁽²⁾
Й	й	y
К	к	k
Л	л	l
М	м	m
Н	н	n
О	о	o
П	п	p
Р	р	r
С	с	s
Т	т	t
У	у	u
Ф	ф	f
Х	х	kh
Ц	ц	ts
Ч	ч	ch
Ш	ш	sh
Щ	щ	shch
Ъ	ъ	" ⁽³⁾
Ы	ы	y ⁽³⁾
Ь	ь	' ⁽³⁾
Э	э	e
Ю	ю	yu
Я	я	ya

German	English
w	v
s	z
ch	kh
tz	ts
tsch	ch
sch	sh
schtsch	shch
ja	ya
ju	yu

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THE SPILITE-KERATOPHYRE FORMATION IN THE REGION OF THE BLYAVA DEPOSIT IN THE URAL MOUNTAINS (PART 1 OF 2)¹

by V. A. Zavaritsky

• translated by Grahame Spragg² •

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Translated from Spilito-keratofirovaya formatsiya okrestnostey mestorozhdeniya blyavy na Urale: Trudy Instituta Geologicheskikh Nauk, Issue 71, Petrograficheskaya Seriya, no. 24, 1946, p. 1-83. Edited by G. C. Amstutz, University of Missouri.

²University of Missouri, School of Mines and Metallurgy, Rolla, Missouri.

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INTRODUCTION

The green volcanic rocks of the Ural mountains are attracting the attention of geologists more and more every year. This can be explained by the fact that there are to be found in them all the Ural pyrite districts that have become world famous. Up to now, however, not much significance has been attached to the petrology of these rocks. When connecting the formation of pyrite deposits with different intrusions, geologists have considered the greenstones only as host rocks.

The supposition, made by A. N. Zavaritsky [1] in 1936, that pyrite deposits and their host rocks could be products of one and the same process, makes us more concerned about these rocks. In each region it is necessary to try to determine the character of the very old volcanism on the basis of detailed petrological analyses. The present study presents the first attempt at such analyses. The volcanic rocks of the western slope of the south Ural mountains, containing the copper-pyrite district of the Blyáva, served as my objective. It is not by accident that these rocks were specifically chosen for the start of my analyses. The only slightly broken layered rocks and the absence of dislocation-metamorphism gives them an advantage over the rocks of the eastern slope, which in most places are so much more strongly metamorphosed, or even converted into metamorphic schists, that it is impossible (in the light of present day knowledge) to clarify the details of the very old volcanism.

The present study consists of three parts. In the first we consider the geology of an explored formation, touching on questions connected not so much with the geological structure of this or that district as with the conditions and forms of the rock itself. Special attention is paid to pillow lavas as generally speaking one of the most original forms of extruded rocks.

In the second part the petrographic features

of the same rocks are described. Here the main questions are the original formation of typical and most widespread rocks, their classification and the elucidation of the mutually reciprocal relations of the numerous varieties.

The third part of the study is devoted to what is called the "spilite problem". By way of a comparison between the rocks under consideration and analogous rocks of other regions we consider questions of their origin.

The material assumed as the basis of this study was collected by me mainly in the summer of 1939. Besides that I have used the results of previous studies in the region of the Blyavinsky neighborhood made in 1934.

The part on the thin sections was kindly given me by P. S. Markov, and I take this opportunity to express my gratitude to him.

I consider it my duty to thank also Zavaritsky and V. S. Sobolev for their advice and interpretation of several questions, and also O. M. Glazov for his evaluations of chemical analyses and his help in the preparation of the illustrations.

SECTION I. THE GEOLOGY OF THE SPILITE-KERATOPHYRE FORMATION ON THE WESTERN SLOPE OF THE SOUTH URAL MOUNTAINS

Chapter 1. General Geology

The extrusive rocks on the west slope of the south Ural mountains have a limited distribution. They are developed only along the Orsk-Orenburg railway and towards the south from there in the direction of Mugodzhär. The limits of their extent are shown on the map (figure 1).

Volcanic rocks are situated here in the middle of lower and middle Palaeozoic sedimentary formations, and join with them into gentle folds. This folded complex lies on Pre-

ambrian metamorphic schists which come out from under it and form the central part of the Urals range, the so called Ural-Tau.

In the West the limit of the region of volcanic rocks, which has received the name of the Sakmarsky zone, borders on the so called Orenburg zone, the region of the limit of upper Palaeozoic sedimentary formations. In the opinion of N. K. Razumovsky, L. S. Librovich and other geologists, the line of a large tectonic disturbance of the thrust type represents the boundary between these areas.

Intrusions of basic and especially ultrabasic rocks in many places break through the folded complex of the Sakmarsky zone. Beside that, the remains of horizontal layers of Cretaceous and Tertiary marine sediments are well preserved here and there on its eroded surface.

The Precambrian schists, quartzites, and gneisses are the oldest rocks in the areas. They are found not only to the east of the region beyond the limit of the volcanic rocks, but also in the middle of the latter in places of anticlinal uplifts.

Silurian sediments, which have been dated on the basis of fossils, lie clearly unconformably on metamorphic schists. The Archaeothyrid limestones of the middle Cambrian in this region do not play an important role in the geological structure, since they are found only in the form of small detached blocks, and their relation to the extrusive rocks found in them is obscure.

Among the lower Silurian formations several limestones stand out. Lower beds are composed of various sandstones with numerous fauna (*Acrotyra*, *Oculus*, *Lingula*, *Orthis*). Higher ones are represented by violet-red and greenish argillaceous shales together with underlying seams of tuffs of a high silica composition. The latter sometimes displace the shales.

Between the sandstones and the schists are seams of limestones with corals and trilobites (*Platylchoe* sp., *Holotrachebis* sp., *Illeenus* sp., and others). The general thickness of all the suites of the lower Silurian period exceeds 1500m.

Above the lower Silurian sediments lies a complex of clastic volcanic rocks of a spilite-keratophyre formation, siliceous shale underneath them, and radiolarian jaspers. At lower levels siliceous shales contain seams of black lustrous shales with upper Silurian graptolites (*Monograptus nudus*, *Monograptus nilsoni*, *Trigraptus* and others). In jaspers of higher levels are found seams of limestone with a lower Devonian fauna (*Carpinsaya conjugula*, *Trypa reticularis*, *Pentamerus* and others).

Siliceous shales, and all seams found in them which have been dated on the basis of fossils, were grouped earlier into a separate suite (Razumovsky) supposedly lying lower than the main mass of extrusive rocks. It is now established that extrusive rocks lie to a large extent directly on lower Silurian sediments, but siliceous shales are always found among the extrusive rocks, occupying the same stratigraphic position at different levels (V. M. Sergiyevsky, 1934 and others). From this it is clear that we must consider the upper Silurian period as the lower limit of the age of the rocks of the spilite-keratophyre formation, and not the lower Devonian, as Razumovsky maintained.

The complex of extrusive rocks and siliceous shales is separated from the lower Silurian period by a stratigraphic unconformity. In many outcrops, for instance at the Troyitsky settlement, it is evident that horizontal layers of spilites lie on steeply dipping (50 to 60°) strata of shales and tuffs of the lower Silurian period. The direction of the folds in the lower Silurian sediments on one hand, and in the younger sediments on the other hand does not differ much. The general thickness of all upper Silurian and lower Devonian suites exceeds 1000 m.

The upper age limit of the complex of extrusive rocks and siliceous shales is obscure. Over the greater part of the area of their occurrence the younger accumulations are eroded away, and only in the north, as Razumovsky points out, does the thickness of the sandstones and shales of the upper Devonian period (the so called Zilairsky-suite) lie discordantly on keratophyres. This mass is graywacke-sandstones and quartz shales. In some rocks pebbles of keratophyres are found. The upper Devonian age of this portion is set according to the finds of plant remains (*Archaeisigillaria primaera*).

With respect to tectonics the region of the spilite-keratophyre formation on the western slope differs strongly from others of the regions of the south Ural mountains. The thickness of the extrusive rocks, the tuffs and the siliceous schists is gathered here into gently sloping cupola-shaped folds. Observed angles of dip fluctuate around 15 to 20° and only on rare occasions do they reach 40 to 50°. Strata of extrusive rocks and tuffs often lie absolutely horizontally. Siliceous schists behave somewhat differently. They are usually gathered into small, steep folds with amplitudes from several decimeters to several meters. The areas formed by the schists are more or less equant. Therefore Razumovsky considering that these rocks always lie beneath spilites, proposed steep cupola-shaped uplifts and an angular unconformity between the shales and the extrusive rocks. However, an exact tracing

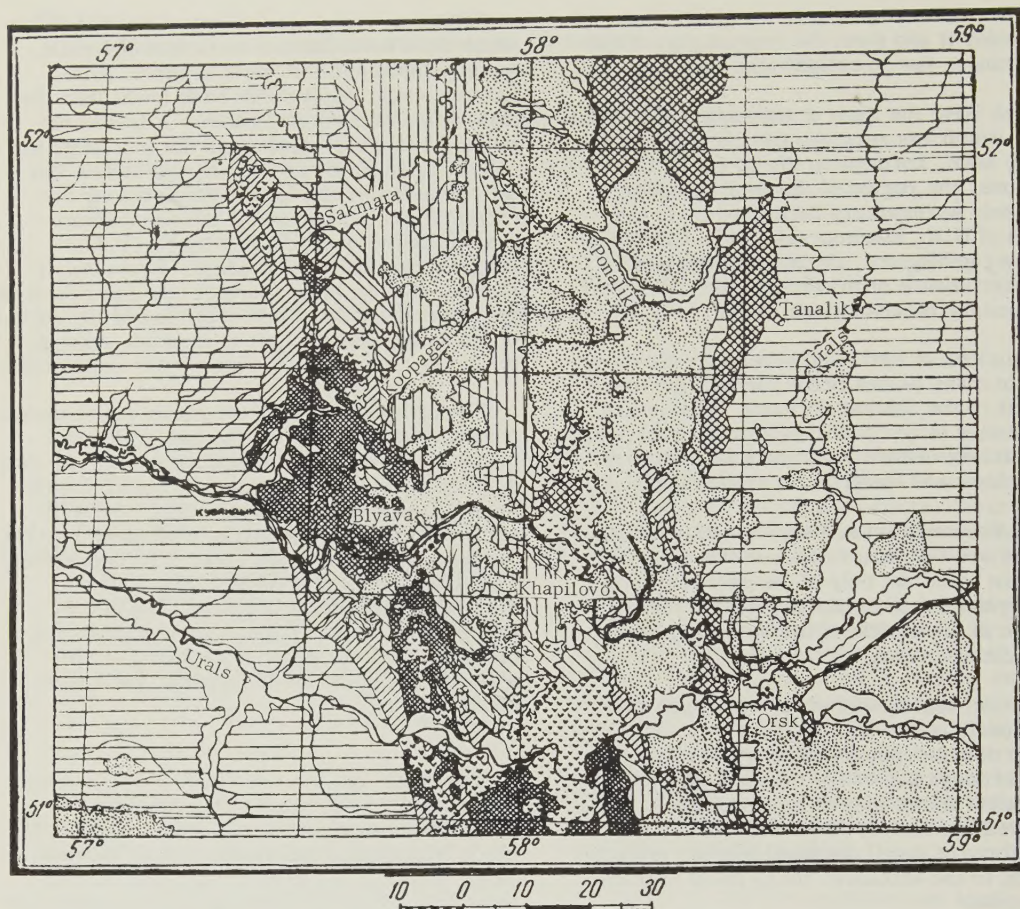











FIGURE 1. Schematic geological map of the south Ural mountains between 51° and 52° N. (A portion copied from "Geological map of the Ural mountains", scale 1:500,000. Pub. G.G.W 1939)

- | | |
|--|--|
|  1 - Precambrian metamorphic complex |  5 - the same rocks on the east slope of the Ural mountains |
|  2 - lower Silurian sediments |  6 - Devonian and upper-Paleozoic sediments |
|  3 - upper Silurian siliceous shales and jaspers connected with the formations that have been studied. |  7 - intrusions of ultrabasic and basic rocks |
|  4 - volcanic rocks of the spilite-keratophyre formation that has been studied |  8 - Mesozoic and Tertiary sediments |
| |  9 - Recent sediments of rivers |

[map] of the outcrops of shales indicated that the boundary between them and the spilites is a function of the relief, since the shales lie slantwise on the spilites (or the other way round). Evidently, disharmonic folding has taken place here.

Fault disturbances in the area under consideration are also apparently insignificant. Small vertical dislocations with amplitudes of several meters are observed in numerous artificial and natural cavities. However, these dislocations do not play a great role in the geological structure. There is insufficient founda-

tion for the view that in the described area there exist steep tectonic breaks with amplitudes of several hundred meters accompanied by zones of fracturing.

Chapter 2. The geological structure of the formation consisting of volcanic rocks

The stratigraphic section and the sequence of eruptions

The stratigraphic sequence of different rocks of the formation under consideration is still not perfectly clear. However, it is certain that the

main mass of basic rocks (spilites) is found at the bottom of the whole sequence, whereas acidic rocks (keratophyres) make up its upper portions. Having an almost horizontal layering, the keratophyres form the crests of heights, and the spilites appear in lower places.

By their outward appearance the keratophyres are so different from the spilites that the division of the whole thickness into a lower series of spilites and an upper series of keratophyres is quite natural. This division was introduced as long ago as 1937 by Razumovsky, and from that time it has been the basis of different tectonic theories. A sharper differentiation between these formations was drawn by A. V. Shabakov (1934), who proposed that even a stratigraphic discrepancy existed between them. However, the observations on the bedding of the different rocks contradict the idea of a sharp break between spilites and keratophyres.

As far back as 1934, at the time of the detailed survey of the environs of the Blyavinsky deposit, we saw that rocks were often found with a composition midway between spilites and keratophyres. Beside that, in 1939 there were observed in a series of places layers of tuffs and agglomerates of a keratophyric composition in typical pillow lavas of spilites. In tuffs it is often possible to encounter separate flows, both of keratophyres and spilites, in which some flows of spilites are found as well as stratigraphically higher layers of keratophyres.

For example, on a summit near the village of 3rd Yumagoozino the flow of the spilite (fig. 2) lies on keratophyric agglomerates, in which portions of keratophyres are found lower on the slope.

Thus, alternations of extrusions of spilites and keratophyres at definite times of the volcanic activity is certain. Therefore, such occurrences are also possible when the spilites lie directly on keratophyres. It is necessary to bear this in mind when setting up different geological sections.

The alternations of spilites and keratophyres indicate that there was no interruption between their extrusions, and consequently there cannot be any question of a stratigraphic break between them. Spilites and keratophyres appear as products of a single volcanic cycle and therefore they must be united into a single spilite-keratophyre formation. The existence of rocks of an intermediate composition confirms the correctness of this conclusion. Besides that, it indicates the fact that the composition of lavas in the course of a whole volcanic cycle changed from spilites into keratophyres more or less continuously.

Extrusive bodies and subdivision of rocks

Extrusive forms in ancient lava beds have always interested geologists, since we may judge from them the character of volcanic eruptions of the remote past. Unfortunately, they are in

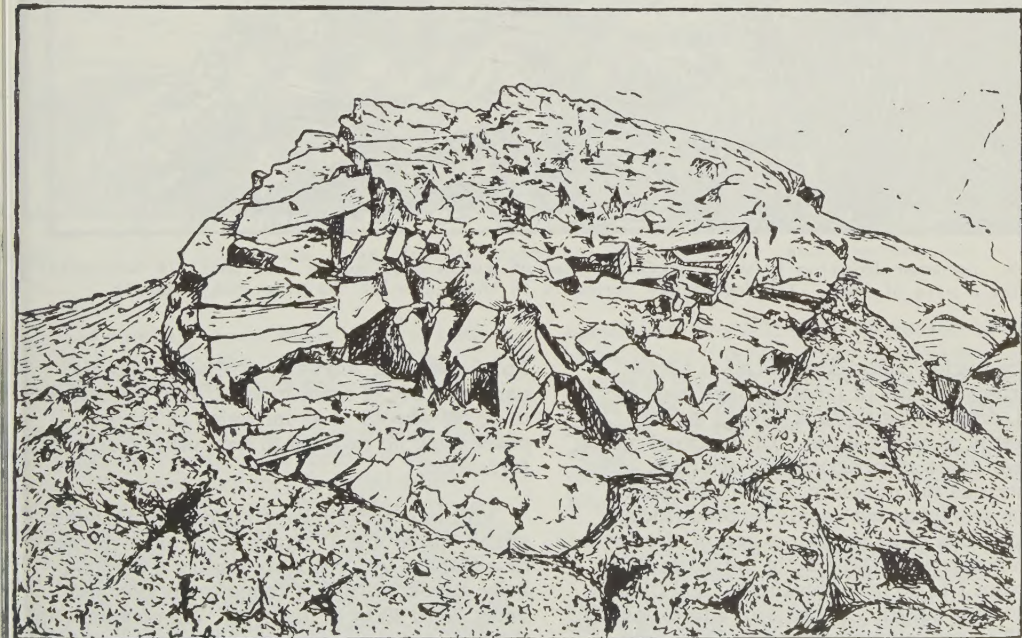


FIGURE 2. The remains of a very old flow of spilite on keratophyric agglomerates, in the neighborhood of the village of 3rd Yumagoozino. One can see the relationship between the arrangement of jointing fissures and the form of the flow.

many places inaccessible for studying. The mass of extrusive rocks, undoubtedly consisting of many beds or flows, are often so deformed in time of folding that all traces of a primary ("volcanic") texture disappear. On the eastern slope of the Ural mountains even the least changed extrusive rocks in outcrops acquire the appearance of schists on account of an intense crushing and cleavage. In the region under consideration, where deformations of the series of volcanic rocks are negligible and cleavage is completely absent, extrusive bodies have been preserved. It is true that conditions of exposure here are unfavorable for observations of these bodies, but with detailed research we may obtain some idea about them.

tion, almost horizontal, and near 20 to 30 cm in diameter, giving the outlets of the dike the shape of a log pile (fig. 3). In steep dikes a parallelopiped-like jointing is observed. Here, more than in the first case, there are pronounced fissures parallel to the sides of the dike (fig. 4).

Dikes of spilites are almost never encountered, being usually hidden under flows. Only near the Troitsky settlement (5 km to the northeast of the Blyavinsky mine), where spilite beds are eroded, are there several of these dikes uncovered in lower Silurian shales.

In other places dikes of rocks have been found greatly resembling diabases, and in 1934



FIGURE 3. Columnar jointing in small vertical dikes of quartz-keratophyre. Columns lie horizontally like a pile of logs. The Novozemsky ravine is in the region of the Blyavinsky deposit.

In the region under consideration almost all extrusive forms of layering have been observed, from dikes to thick covers and agglomerates of pillow lavas. Of the crosscutting features, dikes of keratophyres are most often encountered, the majority of which present roots of overlying but usually eroded flows. The transition of a dike into a bed is often even seen in outcrops. The size of the dikes fluctuate from 1 to 50 m in thickness, and from 5 to 500 m in length. Dikes with a thickness of 15 to 20 m are encountered more often than others. In small dikes columnar jointing can be seen. The columns have the shape of parallelopipeds, extremely elongated in one direc-

I took them for intrusive spilites. After a careful study of them in 1939, they turned out to be off-shoots of an intrusion of gabbro, possessing no relation to the spilite-keratophyre formation.

Volcanic necks have an uncommon secant shape. Near the Troitsky settlement the neck of a diabase was found to display a vertical body almost perfectly round in section and about 30 to 40 m in diameter. In contact with the wall rocks the diabase in this neck changes into an aphanitic rock. The surrounding argillaceous shales are somewhat turned into hornfels at the contact. Necks of keratophyres were not



FIGURE 4. A thick dike of quartz-keratophyre, appearing in spilites. A parallelopiped-like jointing and an array of fissures is seen.

observed, although some short dikes are so thick that one may take the shape of it as transitional between a dike and a neck.

Pillow lavas of spilites are scattered abundantly among the extrusive forms of these rocks. They are examined in detail below. Ordinary flows and covers of spilites are comparatively rare, and their thickness apparently does not exceed 5 to 10 m, in spite of the significant size of every spilitic formation. The jointing in the beds is lumpy or in parallelopipeds. In several beds occurring in pillow lavas near the village Rakityanka columnar jointing is well displayed. Here, columns also processing the shape of elongated parallel-

opipeds are vertical to the horizontal bedding of the cover. In one of the covers revealed in a railway cut, the columns are slightly bent, apparently in the direction of the movement of the lava. On the edge of this flow the columns radiate in a fan shape. Parallelopipedic jointing often changes into "pillow" or concentric shell-like jointing (fig. 5).

The features of this separation are as follows: first, here as also with a parallelopiped jointing, the whole mass of the rock is cut by parallel fissures into pieces of 20 to 30 cm; secondly, concentric shell-like sections get detached from all such pieces in consecutive round breaks, so that a piece takes



FIGURE 5. "Pillow" or concentric-shell [exfoliation] jointing in diabase. Its formation from rectangular jointing can be seen. The road cut is located between the village of Herzonka and the Blyavinsky mine.

on an almost spherical shape. In outcrops, where spilites jut out impressively into the air, the main parallel fissures are not visible, and only spherical bodies are found in a gravel-like [Ed. ?] mass with diameter of 10 to 15 cm with concentric fractures.

The area of distribution and the shape of individual beds are not always clearly recognized. Apparently they do not extend very far and are not comparable to the extent of the large flat cake-shaped pillows of the pillow lavas.

Outflows of spilites were observed still more rarely than beds. The already mentioned flow was well preserved on keratophyric agglomerates at the village of 3rd Yumaguzino. The arrangement of the fissures of the blocky jointing emphasizes its shape (fig. 2). It is impossible to establish the length of it, but it undoubtedly exceeds 10 to 15 m. It is more difficult to make out separate flows in a mass of keratophyres, since in the keratophyres there are no visible changes of structure in the external parts of extrusive bodies. Where they can be distinguished, for example the alternation with layers of tuffs, their thickness was measured as 1 to 5 m. Their jointing is lumpy but rectangular.

Pillow lava of spilites and the process of their formation

As has already been said, a large part of the spilites of the formation being described is found in the form of the so-called globular or pillow lavas. In natural outcrops which often

display only road ballast material, pillow structure is not apparent because of the weathering of the rocks. However, in all vertical artificial sections it is perfectly well visible.

The shapes of the pillow lavas of the spilites described in this paper do not differ from the majority of those described in the literature. The whole mass of lava here consists of detached, small ellipsoidal bodies (pillows), lying one on the other so that an upper one apparently partially covers one or two of the underlying ones. Usually there were no spaces between the pillows, but sometimes there are found between them in a strikingly odd way crushed siliceous schists or a tuffaceous material. In the road cut at the village of 3rd Yumaguzino separate pillows are completely surrounded by a heavily weathered tuffaceous material. The size of the pillows varies from 0.1 to 0.2 m [? 2 m Ed.] in diameter. Pillows are seen most commonly from 0.4 to 0.7 m. The shape of the pillows greatly depends on their size: large pillows have an irregular, sack-shape, or round-loaf-like shape (fig. 6), but small ones have the shape of almost perfect spheres.

The peripheries of the pillows always consist of a less crystalline rock. Besides that, they commonly contain more amygdules which are often arranged in concentric zones. In several places the periphery of pillows is not only chilled, but also greatly enriched with oxides of iron. It has the appearance of a red, very compact crust with a thickness of 2 to 5 cm.

Every pillow has, it would seem, its own



FIGURE 6. Pillow lava of spilites in the neighborhood of the village of Rakityanka. The shape of the pillows is most typical. Their detached nature can be seen.

system of fissures of jointing. In the periphery the fissures are always parallel to the outer surface of the pillow, but nearer to the center they are more or less perpendicular to it. The body of the pillow lava on the whole cannot represent an irregular conglomeration of pillows. It consists of several layers up to 10 m thick, in each of which pillows of a more or less identical size prevail. In the railway cut near the station of Blyava these beds which dip here at an angle of 20 to 25° are easily seen (fig. 7). Sometimes between these beds there are seams of black siliceous shale or red radiolarian jaspers of a thickness between 0.5 to 1.0 m. The seams and also the presence of shale between the pillows indicate the underwater conditions of the extrusions of spilites here described.



FIGURE 7. Pillow lavas in the railway cut at 232 km near Blyava halt. The gentle dip of the whole layer of the pillow lavas can be seen. The pillows consist of the microlitic variety of spilite shown in figure 13a.

The pillow lavas have the unusual shape of bedding of extrusive rocks, and the mode of their formation appears to be the object of a discussion which has already been carried on for several decades. Pillow lavas are very common in spilites. There has even existed a tendency to use the terms "spilite" and "pillow lava" as synonyms, and only examples of pillow lavas of another composition have compelled us to give up this idea.

The term "pillow lava" is still not used by everyone in only one sense. The majority of researchers use the term only for those in which the separation of pillows has taken place, undoubtedly when the lava was still fluid. However, some use the term generally for lavas which in outcrops look as if they consisted of separate spherical or ellipsoidal bodies irrespective of their origin.

In the Russian literature, particularly in the publications of F. Yu. Levinson-Lessing, pillow structures are thought to have been formed without exception in the form of jointing. It is impossible to consider this correct since jointing in the conventional sense is a division of the mass of a rock by a system of contraction due to the cooling of the already hardened rock. But in the majority of pillow lavas and in those described at that date by Levinson-Lessing, the division into pillows undoubtedly had taken place before the hardening of the lava. We find irrefutable evidence in the features of the

shape and inner structure of individual pillows, and also in the presence of a sedimentary material between them. Thus the process of formation of pillow lava is completely different from the forming of "globular separations" [i. e. spheroidally weathered rocks. Ed.] described above (fig. 5).

It is still not quite clear how the detachment of the pillows took place. In the opinion of the English geologists (C. Reid, H. Dewey, J. S. Flett and others), pillow lavas formed into under-water, broken-up extrusions, when lava, having flowed out onto the sea sediments, broke up (from the action of the water on it) into separate spheroidal drops surrounded by a casing of steam. Later on, on account of the rapid formation of a hard crust, these spheres became practically hard and rolled along the sea bed. The advancing flow of lava in such a case had to consist of an agglomeration of such rolling globules.

Another explanation is offered by the American geologist J. V. Lewis [4]. In his opinion the flow of lava from the start did not have a pillow structure. Later on, when its upper and already hardened crust cracked, and under the pressure of rising lava thin streams of it percolated through into the fissures that had formed, round-loaf-shaped bodies appeared out of the liquid lava with a hard but elastic casing. New portions of lava enlarged these bodies, their covers cracked with hardening, and on their surface new bodies were formed. As a result the whole mass of the lava turned into a number of spheroidal bodies joined to each other with thin necks. However, in the opinion of J. V. Lewis, it is impossible to conceive that the formation of these bodies took place without a relative dislocation of them. In such a case a rupture of the necks and a full isolation of the pillows is inevitable.

These two explanations of the formation of pillow lavas appear up to the present time to be the most widespread ones in petrology. But several other explanations still exist. Recently J. T. Stark [6] offered his opinion that gases contained in the lava play the main role in the formation of a pillow structure. Just like Lewis, he considered that the flow of lava had no pillow structure at the beginning. Pillows began to form later in the central parts of the flow as a result of rotatory movements of accumulated gases there. The lavas from the islands of Tovarishchestvo, described by Stark, differ from typical globular lavas, and therefore his explanation, as H. E. McKinstry has already noted, ought not to extend to the latter.

The shortcoming of Lewis' theory, i. e. the so-called "bulbous budding theory", appears to be as A. Wells [41] noted, the denial of water as an agent in the formation of globular lavas. But typical pillow lavas were undoubtedly formed

under water, as shown by the presence of a sedimentary material among the pillows, sometimes even with sea fauna. Besides that, T. Anderson [3] observed at first hand at the time of the eruption of the recent volcano Matavanu (o-v Savayi [Is]), that the stream of ordinary ropy lava flowing into the water began to acquire a pillow structure greatly resembling that observed in typical pillow lavas. The theory of Lewis has other shortcomings too. For example in typical pillow lavas there are no traces of connections between their necks of the kind that existed among the separate pillows. In those cases, where pillow lavas are formed of pillows with an overall diameter of 10 to 30 cm, it is impossible to imagine that the latter formed out of each other. They are without doubt absolutely independent bodies that have been formed at the same time. The independence of the pillows is also obvious in those cases where they are completely surrounded with sedimentary material.

Thus the theory of the English geologists, which proposes that the pillows became isolated immediately as the lava appeared on the surface, agrees more than any other with the facts observed in typical pillow lavas.

In the opinion of Wells [41] the shortcoming of this theory is that there is some mysticism in the explanation of the reasons and methods of the motion of the lava flow. In his words, the inclination of the sea floor is so slight that the half hardened globules would scarcely have been able to roll along it. It is impossible to agree with this observation of Wells. The dislocation of the individual pillows followed at once after their isolation. At least they piled up on each other as long as they were very soft, probably when they were still "drops" of molten lava. Besides that, there are no signs that they really rolled, i. e. they had rotatory movement.

To some extent it is possible to compare the condition of the lava at that moment with emulsion being formed with the rapid mixing of a heavy liquid (for example, bromoform) and water. If a cover of solid lava had been able to flow over this floor, it is all the more possible for a more mobile mixture of lava, water and gas to spread over it.

At the beginning K. Reid and H. Dewey [23] proposed that the dislocation of the pillows was furthered by a decrease of their individual weight on account of the forming of numerous bubbles in the lava. However, pillow lavas with large quantities of bubbles are rare, and this contradicts such an explanation W. N. Benson [50] added to the proposition of the English geologists by admitting the possibility of the formation of a pillow structure in lavas intruding into wet sea silt. This admission does a better job of explaining the presence of the sedimentary material among the pillows.

however, it is impossible to agree with the observation of Benson that a pillow structure does not point to an extrusive origin of rocks. The intrusion of lava into sea silt does not differ essentially from its usual extension.

The main feature of a single flow of pillow lavas is the fact that its dimensions are very indeterminate. It is hard to establish what number of individual pillows appeared on the surface after a single eruption. In the Blyatsky region one sometimes encounters a certain separation between individual beds, and it is possible to admit that the pillows of a single layer were formed simultaneously. In such a case the extrusions of lava were not very large. The thickness of a layer is 10 to 20 m; that is very sizeable in comparison with the thickness of the whole spilites series.

The differences in size and shape of the flows apparently depend on the conditions of effusion. With a rapid mixture of lava and very small globular drops of lava were formed, while with a slower mixture they were naturally larger. This is not hard to understand if we again make an analogy with emulsions. Lavas, possessing a pillow structure, were also formed in one and the same way. The considerations brought out above are attributed only to the most typical pillow lava, like the well known lavas in England, which apparently are identical with that of the spilites of the formation being described.

The clastic-volcanic deposits

Until most recent times clastic-volcanic rocks have been considered comparatively rare among the rocks of the formation being considered. Typical agglomerations ("tuffaceous breccias") of a keratophyric composition were found as long ago as 1934, at the time of the detailed survey of the neighborhood of the Lavinsky deposit. At that time the character of distribution of these rocks was still not clear, and only in 1939 after a survey of a larger area was it found out that there are not fewer keratophyric agglomerations than lavas of the same composition. Sometimes the agglomerates even predominated significantly over the lavas, and the latter was found in them in the form of separate small flows. The spread of keratophyre agglomerates indicates that eruptions of keratophyres were accompanied by violent explosions and were not peaceful fissure effusions as has been supposed up till now.

The composition of the clastic-volcanic deposits differs according to their stratigraphic position. Agglomerates of a spilitic composition are encountered in the mass of the spilites. These rocks are only exposed in two places

1) in the railway cut near the village Alchembayevo and 2) at the confluence of the Baka and Kooragan rivers. Near the village Alchembayevo they change into agglutinates ("lava breccias", i. e. rocks in which cement is not tuffaceous, but variolitic). In upper horizons of the spilitic series, and also in between the spilites and the keratophyres agglomerations of a mixed [hybrid] composition was found. In them together with predominant fragments of keratophyres and near them hornblende porphyrites are fragments of most diverse rocks: spilites, siliceous schists, lower Silurian felsites, and red argillaceous schists. The latter are encountered especially often. This allows [us] to assume that volcanic hearths were found in that part of the earth's crust which consisted of just these schists, i. e. apparently at a comparatively shallow depth (1 to 1.5 km).

The relative numbers of xenoliths decreases quickly in the upper beds. Fragments of keratophyres and hornblende porphyrite gradually give place to fragments of quartziferous keratophyres. The tuffaceous cement is usually of a more basic composition.

The size of the xenoliths varies greatly, but fragments of 5-10 cm in diameter predominate. Sometimes lumps are found up to 0.5 to 1 m. The larger fragments have an angular shape, and the small ones are in great part rounded. The relative amount of fragments and cement are different in different layers. Usually the fragments comprise two thirds of the whole mass of the agglomerate. In some places there are less of them and the agglomerates gradually change into tuffs.

The schistosity in the tuffs and agglomerates is not noticeable either in specimens or even in large blocks; but in outcrops it is clearly visible. Sometimes it is emphasized by thin (0.1 to 0.3 m) streaks of tuffites.

Agglomerates and tuffs are rarely encountered in a fresh state. In eroded sections half-loose rocks with protruding hard fragments of keratophyres are seen. At first glance their outcrops differ little from outcrops of spilites, and probably, therefore vast areas made up of agglomerates on geological maps, are painted over with the colours of spilites. During the weathering of the agglomerates only the latter and keratophyres are uncovered. Therefore the agglomerates were completely left out during the mapping, and the quantity of keratophyres was exaggerated. Less evenness on eroded surfaces of agglomerates is visible especially on hills with steep slopes, underlain by horizontal alternating layers of these rocks and of keratophyres. Here the separate flows of keratophyres form protruding steps that are located one above the other.

SECTION II. THE MOUNTAINOUS BELT WITH THE SPILITE-KERATOPHYRE FORMATION OF THE WESTERN SLOPE OF THE SOUTH URAL MOUNTAINS

Chapter 3. Spilites

Among the basic rocks of the formation under consideration are encountered both true spilites, characterized by the presence of a fully albitized plagioclase and ordinary diabases and other extrusive rocks of a basaltic composition. It is impossible to separate these rocks, since the sections with albitized and with non-albitized plagioclase are encountered not only in one lava bed, but sometimes also in random specimens. Diabases with basic plagioclase are comparatively rare, and therefore every group is considered under the general name spilite.

When describing spilites, it is not suitable to hold strictly to an ordinary petrographic classification and to put under the same heading rocks with an identical degree of crystallinity. In a given case this grouping would be formalistic, not reflecting natural geological associations. For example, it is more natural to unite variolites, forming the peripheral parts of lava beds, with those rocks of greater crystallinity from the central parts of the same lava beds, rather than with other variolites forming independent beds.

A comparison of the petrography of spilites with data about the geological conditions of their bedding allowed us to establish types whose different features do not depend on the degree of crystallinity of the rocks.

Diabases from the hanging wall of the Blyavinsky deposit and the spilites connected with them

The features of the spilites of this type appear to be 1) the presence of faintly tinted brownish pyroxene, 2) a faint trace of albitization of plagioclase in the more crystalline varieties, 3) a strong elongation of plagioclase laths in less crystalline varieties, and 4) the more or less prismatic shape of pyroxene grains.

These rocks are not widely scattered. They occupy a significant area only near the Blyavinsky deposit, to the west and northwest of it. There are hardly any other spilites here. Other than in the region of the deposit, the spilites of this type are encountered at the village of the 3rd Yoomaguzino. But isolated specimens of them are known also in other places.

They differ according to the degree of their crystallinity: a) medium grained diabases, found more often with basic plagioclase and more rarely with albite, b) small-granular

diabases, usually with albite, c) spilites with a hyalo-ophitic texture, d) variolitic aphanites, and e) microlitic varieties, often with basic plagioclase.

Medium grained diabases are less widespread than other varieties, but they are well known among the rocks of the hanging wall of the deposit. As could be seen in several places they are restricted to the central parts of the lava beds, the thickness of which exceeds 3 to 4 m. The transition into finer grained rocks, composing a large volume of the lava beds, is a gradual one. Fine grained diabases form beds with a thickness of not more than 3 m, or are encountered in the form of pillow lavas, where pillows are not less than 1.5 m in diameter. The peripheral and less sizeable parts of these pillows or beds are formed by aphanitic varieties: spilites with a hyalo-ophitic texture or variolitic aphanites. These varieties are not only encountered in the periphery of the beds, - a large part of typical pillow lavas to the northwest of the deposit is formed exclusively from spilites with a hyalo-ophitic texture. The dimensions of the pillows here are smaller (0.7 to 1.0 m). Variolitic aphanites make up a comparatively small part of the spilites under consideration and are situated only in the same peripheral parts of extrusive bodies. The transition to other varieties is gradual.

Microlitic varieties change directly into diabases or spilites with a hyalo-ophitic texture. They form small beds and globular lavas to the southwest of the deposit. Pillow lavas differ slightly from typical less detached pillows. Sometimes they bear a resemblance to flows of rhyolite lavas.

Among the diabases that form thick beds a concentric shell-like separation can be observed (fig. 5).

The rocks being described stand out noticeably among the other types of spilites of this region by their external appearances. They are significantly darker, being almost perfectly black. A greenish tint, more noticeable in aphanitic varieties, is observed on fresh specimens of drill holes and adits. A brownish of rusty tint predominates on erosion surfaces. Granularity is distinctly visible in diabases, thanks to the conspicuous brilliance of cleavage planes of the laths of darkish plagioclase from 1 to 3 mm in length. In a fresh break of aphanitic varieties some fibrous texture is visible, which, as we will see below, characterizes the microstructure of the rock.

The less crystalline varieties of spilites in the periphery of the beds sometimes change into amygdaloidal rocks. However, more often only separate amydules filled with chlorite or calcite are observed. Rocks with a more or

ess marked porphyritic texture were not encountered.

Mineralogical composition: The composition of spilites of the type under consideration is the same in respect to the features of individual minerals, but varies according to their quantitative ratios.

The main components appear to be: 1) plagioclase, represented both by the albite normal or spilites, and also by preserved primary basic plagioclase, 2) monoclinic pyroxene possessing an ever visible brownish coloring, and 3) glass, completely decomposed into a finely-crystalline aggregate of chloritic minerals. Accessory magnetite is sometimes quite abundant.

Plagioclase: As already noted, primary basic plagioclase is characteristic for medium grained diabases, although it is also sometimes encountered in other varieties. In microlitic spilites it can be determined only according to the index of refraction, clearly higher than that of Canada balsam.

In diabases basic plagioclase looks completely fresh. Saussuritization or prehnitization is not typical of it.

The composition of the plagioclase is sufficiently constant, and corresponds to the composition of basic labradorite (55 to 70 percent An). Sometimes plagioclase is somewhat zoned, and in a very narrow peripheral zone the content of anorthite molecule goes down to 40 to 45 percent, judging by the angle of extinction in the sections $\perp [100]$, which changes from 30 to 35° in the center, to 22 to 25° in the periphery.

Laths of labradorite possess an elongated prismatic shape parallel to $[100]$. Twins are usually simple, according to the albite - Carlsbad law: $B \wedge Z = 74$ to 78° , $B \wedge Y = 32$ to 47° , $B \wedge X = 63$ to 46° , $B = \perp [001]$ in the plane of (010). Polysynthetic twins were observed more rarely.

Labradorite is found not to be completely replaced by albite in fine grained diabases. The boundary between albite and labradorite is always sharp. Albite develops in labradorite in the form of irregular patches (fig. 8). Albite can be well recognized by the following properties: 1) polysynthetic twins, 2) the maximum angle of symmetrical extinction in these twins, equal to 15 to 17°, 3) the mean index of refraction is 1.53^{\pm} and 4) birefringence near 0.010. These data do not prevent us from con-

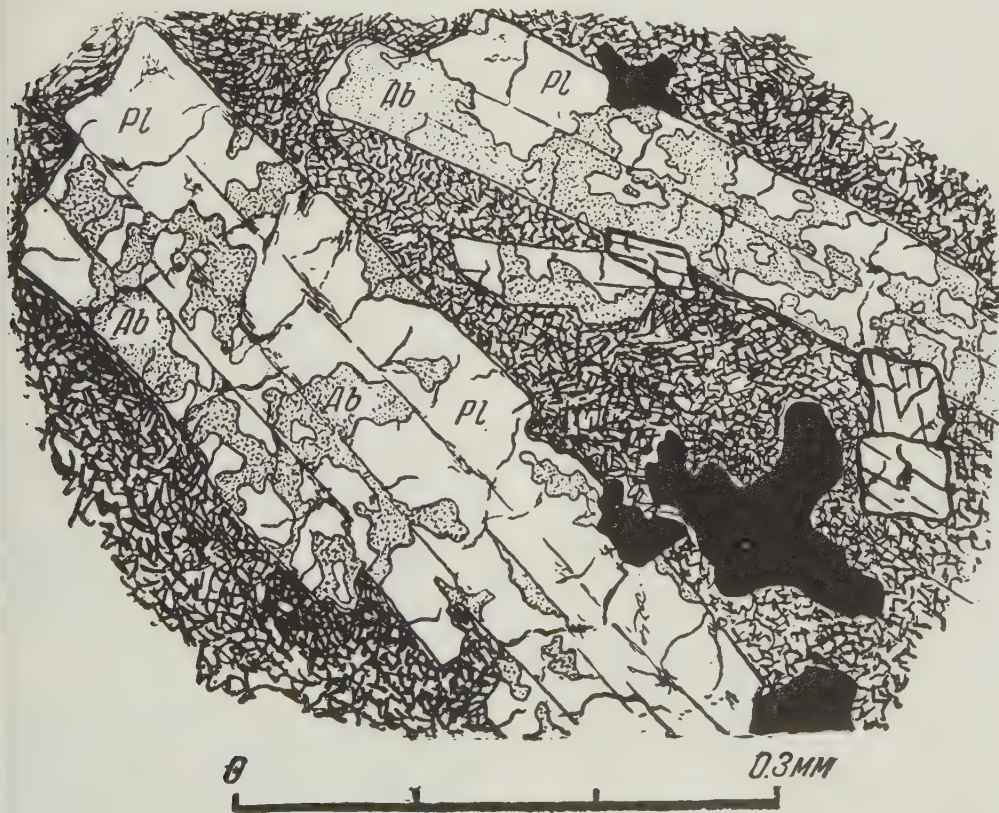


FIGURE 8: Plagioclase (Pl) not completely replaced by albite (Ab) in fine grained diabases from the region of the Blyavinsky deposit. A typical example of metasomatism. The scale at the bottom is in millimeters (the same applies to all figures after this).

fusing it with zeolites. Polysynthetic twins are however rare and albite is usually not twinned.

It is interesting that a specimen with not completely albitized plagioclase was taken out of a bed with a thickness of 1.5 to 1.7 m, which stands out distinctly among the pillow lavas containing it, because of a very pronounced vertical columnar separation.

The albite, representing plagioclase in a large part of the spilites here described, also attracts attention by its freshness. It is perfectly transparent, and only sometimes does argillization, typical for secondary albite, appear as spots in larger grains. In very few thin sections albite shows inclusions of secondary calcium minerals: they are either tabular prehnite grains or small grains of zoisite. In these cases albite is more argillized and speckled with scales of sericite.

The content of anorthite molecule in albite, judging from the angles of extinction in sections \perp [100] that are well determined from the crystal outlines, does not exceed 10 percent, which is confirmed by an analysis of several twins. In more widespread albite-Carlsbad twins $B \wedge Z = 86^\circ$, $B \wedge Y = 80$ to 84° and $B \wedge X = 6$ to 10° . In albite twins $B \wedge Z = 12$ to 17° in the presence of coinciding X. The mean index of refraction, tested in several cases by immersion, is 1.528 to 1.531.

The shape of the albite laths depends on the degree of crystallinity of the rocks. The laths are always elongated parallel to [100] and, what is especially characteristic, have inclusions of decomposed glass in the nucleus, sometimes with small grains of pyroxene.

Monoclinic pyroxene is distinguished by its brownish coloring, sometimes with a violet shade and weak pleochroism. The optical properties of this mineral, determined in thin sections of different rocks, are almost identical. The indices of refraction are comparatively high: $\gamma = 1.730$, $\beta = 1.711$ and $\alpha = 1.708$. According to several measurements the birefringence is not large: $\gamma - \alpha = 0.020 - 0.023$. $2V = +50^\circ$, but varies from $+47$ to $+52^\circ$. The dispersion is strong, $\rho > \nu$. The angle of extinction $C \wedge Z$ does not exceed 40° , but it sometimes unexpectedly reaches 36° .

Comparing the optical properties of our pyroxene with the data of the new diagram of A. Winchell [162], it is possible to see that with regard to the angle $2V$, and the indices of refraction, it corresponds closely to pyroxene, the ratio of which is $\text{FeO} : \text{MgO} = 1$, or in molecular percentages, $di = 20$, $en = 30$, $hy = 30$ and $hd = 20$. However, with regard to the angle $C \wedge Z$ there is a significant divergence with the data of the diagram, since ac-

cording to Winchell, pyroxene of such a composition has this angle of not less than 44° . We get the same result when we use the diagram of T. Tomita [163]: according to $2V$ and the indices of refraction, the composition of our pyroxene in weight percentages is: $\text{MgSiO}_3 = 35$ percent, $\text{CaSiO}_3 = 30$ percent and $\text{FeSiO}_3 = 35$ percent. The discrepancy cannot be explained by the influence of admixtures of Al_2O_3 and Fe_2O_3 , since it is known that these mixtures, on the contrary, increase the size of the angle of extinction.

A small angle of extinction, as also a small angle of $2V$, indicates rather that our pyroxene is not rich in Al_2O_3 and cannot be called augite.

The violet color apparently testifies to the presence of TiO_2 .

The shape of the grains of pyroxene is irregular, but is always more or less elongated parallel to [001].

In less crystalline rocks the pyroxene is sometimes found in very elongated prismatic crystals or in needle-shaped groups of micro-lites.

Decomposed glass is present in all varieties depending on the degree of crystallinity. This is a very fine scaly aggregate of chloritic minerals. The scales are usually of submicroscopic dimensions. The different scales are arranged regularly, forming either a spherulitic or concentric-zonal aggregate structure. Spherulitic formations themselves often possess concentric zoning. Several zones in them are replaced by quartz.

The determination of the optical characteristics of chloritic minerals replacing the glass is impeded by the very small dimensions of the scales. We have succeeded in determining that in green varieties $\beta = 1.59$ to 1.60 , $\gamma - \alpha = 0.005$; in yellow-brown varieties β increases in size up to 1.63 , and birefringence, up to 0.010 . The sign of the main zone is always positive, i.e. these chlorites are probably optically negative. Such characteristics allow us to attribute our chlorites to delessites, i.e. to magnesium-ferrous chlorites poor in Al_2O_3 . Comparing these characteristics with the data of A. Winchell's diagram [164], we see that the molecular ratio $\text{Al} : \text{Mg}$ in them is about $20 : 80$, and $\text{Fe} : \text{Mg}$ changes from $20 : 80$ to $60 : 40$.

Together with the usual chlorite a mineral is sometimes found resembling it, with a high birefringence (up to 0.20), positive elongation, and a high index of refraction ($\beta = 1.65 \pm$). It is apparently a viridite, about which little is still known. It is always found in the form of spherulites and accompanied by quartz.

Pseudomorphs after olivine are encountered fairly often especially in the least crystalline varieties. They are recognized by their outlines, which sometimes completely correspond to the outlines of the typical dimensions of prismatic crystals of olivine with the faces (21). Often the shape of pseudomorphs are more or less rounded or irregular on account of the typical "embayments" inherent in crystals formed from melts (fig. 11a).

Usually pseudomorphs consist of a carbonate and a small quantity of serpentinitic chlorite matter. The carbonate, where it has been successfully determined, proved to be calcite ($\rho = 1.658$). Several pseudomorphs consist of antigorite ($\beta = 1.561$ and a positive elongation). These pseudomorphs possess a common cottonhole structure.

Biotite was encountered in several diabases and is usually almost completely converted into chlorite.

Magnetite is found both in large grains, and also in accumulations of small grains in decomposed glass. In large grains it changes to leucoxene, i.e. it is apparently titanomagnetite. Apatite in the form of sparse needles is found only in medium grained diabases.

Secondary minerals, besides chlorite, are encountered in a very small amount. The most common of them are: prehnite, zoisite, epidote, quartz, calcite, actinolite, and iron hydroxide.

Texture and features of the separate varieties: As has already been said, several varieties stand out among the spilites herein described. We will now discuss their features in more detail.

Medium grained: Diabases almost always possess a tholeiitic texture, since there exists in them decomposed glass, filling the angular spaces between the small laths of plagioclase where they are not occupied by grains of pyroxene. The relative quantity of glass varies from 10 to 40 percent. In the glass there are often visible amygdules or microlites and bunches of fibres of plagioclase together with little needles of pyroxene. A main feature of the texture lies in the character of the reciprocal relations of plagioclase and pyroxene. Ophitic relationships common for diabases in the rocks herein described are rarely observed. Their texture is evidently characterized by a marked tendency of the grains of pyroxene toward pseudomorphism (fig. 9a). Pyroxene is found between the laths of the plagioclase, but its more or less prismatic grains often cut into the latter or even cut through them. According to the lessening of the quantity of glass, the texture does not approach a typical diabasic (ophitic), but intergranular or doleritic nature.

The sizes of the grains are more or less identical. The laths of plagioclase are about 2 mm long and 0.1 to 0.3 mm wide. In large grained rocks the length of the laths reach 5 mm with a width of 0.5 mm. The grains of pyroxene are less (0.8 to 1.0 mm in length, 0.3 to 0.5 mm in diameter).

2. Small-grained diabases - differ from medium-grained only in the dimensions of the grains (fig. 9b). The laths of plagioclase are here 0.8 to 1.0 mm long and 0.1 x 0.05 mm wide. The pyroxene grains measure 0.2 x 0.5 mm. The relative quantities of minerals and the character of the reciprocal ratios of pyroxene and the plagioclase here are the same as those of the rocks described above.

Some fine grained diabases possess a texture as it were transitional to the textures of the spilites considered below. In these diabases the quantity of glass increases and the dimensions of the grains are variable. Among the small laths of plagioclase that are a little more elongated than is usual in diabases, separate laths are found up to 2 to 3 mm long. The pyroxene sometimes forms needles (1.0 x 0.2 mm) or parallel groups of several such needles. We may consider several textures of diabases as transitional to textures of variolitic aphanites examined below. The characteristic trait of these textures appears to be the fact that between the plagioclase, chloritic aggregates are not found, but rather a mass of panicked or, more often, cone-like bunches of fibres of plagioclase. Some of these bunches change directly into laths, and therefore the ends of the latter look splintered.

3. Spilites with a hyalo-ophitic texture are more widely spread than the other varieties. The main feature of their texture appears to be solely the elongated shape of the crystals of plagioclase (fig. 10a), the length of which reaches 1.5 to 2 mm with an overall width of 0.05 to 0.08 mm. The wavy flow line in which the crystals are lined up is characteristic. The amount of glass included in them is more than in the laths of plagioclase in the diabases. There were observed even skeleton forms of crystals, the cross-sections of which look like the O and C of square print.

The arrangement of the crystals is disorderedly, but sometimes they form sheaf-like, panicked, radial-radiant groups or are arranged parallel. The decomposition of the glass in these spilites is significantly greater than in the diabases. In places the crystals of plagioclase do not even come into contact with each other.

There is not much pyroxene, and therefore the typical texture is called hyalo-ophitic. This term does not quite convey the features of the texture, but it is the most suitable of the accepted terms.



FIGURE 9. a) Medium grained diabase from the hanging wall of the Blyavinsky deposit; bore-hole 15, depth 52 m (see analysis, Table 1). Labradorite, grains of pyroxene with clear outlines, magnetite, and chloritic mesostasis. A typical intergranular texture. b) Small-grained diabase of the same type. Plagioclase: albite and elongated grains of pyroxene. In the figure can be seen the similarity in structure of medium grained and small-grained varieties.

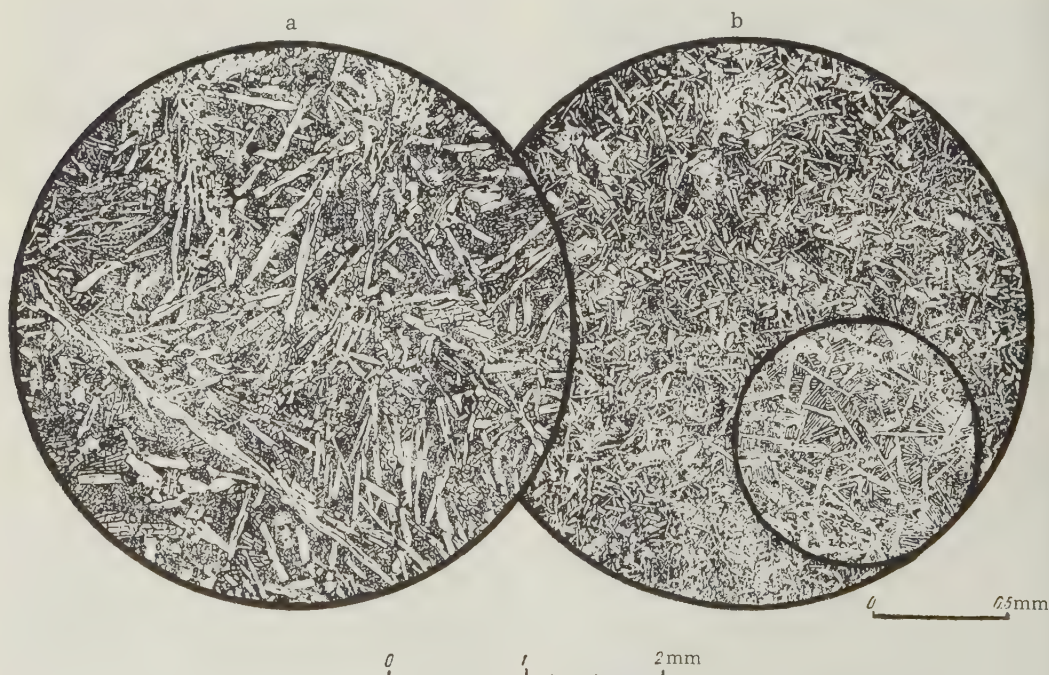


FIGURE 10. a) A spilite from the neighborhood of the Blyavinsky deposit. A typical texture of the rock. Exceptionally elongated laths of plagioclase (albite), small pyroxene grains with clear relief, and a chloritic matrix. b) A microlitic variety from neighborhood of the village of Herzonka. Labradorite-plagioclase. In the small circle is shown the growth of pyroxene needles on microlitic plagioclase.

In the rocks with the typical texture, pyroxene is found between crystals of plagioclase in separate prismatic, but not very elongated, small grains with a size of 0.2 x 0.3 mm. Larger prismatic grains or needles with a length of up to 1.0 to 1.5 mm are comparatively rare. The tendency towards idiomorphism with pyroxene is still more pronounced in the spilites than in the diabases. In some rocks the texture is appreciably different from the typical texture, although it does also have a lot in common with it. Here there is less glass and correspondingly more plagioclase and pyroxene. The spaces between such elongated crystals of plagioclase are completely occupied with pyroxene, which is found in the form of parallel, accreted needles and other similar skeleton shapes. In great part the needles are arranged parallel to the crystals of plagioclase, but they also often grow on them in the form of brushes. Such a texture resembles that of a bird's feather, described by Levinson-Lessing in Caucasus diabases. In the places where the crystals meet, needles of pyroxene growing perpendicularly on them cross each other, forming original skeleton shapes in the form of trellis work.

Variolitic aphanites appear to be the most crystalline variety. Their texture is characterized by skeleton shapes of crystals. The rocks are formed mainly out of decomposed glass, radial or sheaf-like groups of fibers of plagioclase, and the finest needles found between them are almost sub-microscopic grains of pyroxene.

Several paniced formations, distinguished by their brownish tint, consist of sub-microscopic particles of an indeterminate mineral, perhaps also of pyroxene. The size of the "panicles" does not usually exceed 1 mm (fig. 11a), but sometimes they reach 2 to 5 mm (fig. 11b).

5. A microlitic variety appears to be rare among the spilites here described. Its texture is interesting, but it differs from that of the typical basalts in that the pyroxene is encountered not in the form of small grains, but in the form of groups of the finest needles (fig. 10b). This feature and also the flow texture of the microlites give this variety a resemblance to common hyalo-ophitic spilites.

Chemical composition: One can get an idea of the chemical composition of the spilites being described from the analyses of Table 1.

From these data it is clear that despite the unquestionable relationship and continuity of the transitions between diabases and spilites with a hyalo-ophitic texture, the chemical composition of these rocks is completely different.

In comparison with the average composition of basic extrusive rocks the composition of diabase is characterized by a very high ratio of feldspathic calcium to alkali i.e. resembling the composition of calcium-basalt. The chemical composition of a variety with a hyalo-ophitic texture is almost identical to the average composition of spilites in general. Both in dia-



FIGURE 11. Variolitic aphanites from the periphery of flows or pillows of pillow lava. From the neighborhood of the Blyavinsky deposit. a) A most typical variety. Pseudomorphs of calcite can be seen in corroded olivine crystals. Typical paniced formations of submicroscopic particles of pyroxene (?). b) These same formations, growing on thin laths of plagioclase (albite).

base and also in spilite a high content of TiO_2 is striking.

A microlitic variety with basic plagioclase is near a diabase in chemical composition.

A comparison of the cited analyses with other analyses of spilites is made below.

Diabases and spilites of the normal type

Under this heading we will consider rocks forming the main bulk of the spilites on the western slope of the south Ural mountains. They are noticeably different from the rocks described above. The pyroxene in them no longer has a brownish tint and it is found in equant grains, and primary basic plagioclase is hardly encountered here. The laths of plagioclase are not as elongated as in previous rocks. Microlitic varieties are very rare among these rocks. They differ from the diabases only in the sizes of the grains. Here there are none of the important differences in shape and no combination of minerals of the kind observed in previous rocks. Therefore it is possible to distinguish the following only under certain circumstances: 1) Medium grained diabases, 2) small-grained diabases and, 3) microlitic

varieties. The latter sometimes change into rocks of a porphyritic texture in which the phenocrysts consist of pyroxene. In microlitic spilites amygdules are often present, and they possess the shape of real amygdaloidal rocks.

The fine grained diabases are the most widespread. They make up the thick (more than 2 m) covers (at the village of Novaya Rakityanka) or they form typical pillow lava (at 3rd Yumaguzino). The periphery of the blows and pillows of the lavas are always formed of microlitic varieties.

The main mass of the microlitic spilites is found in the form of pillow lavas, an excellent example of which may be seen in the railway cut at 232 km (fig. 7). The size of the pillows here is sometimes measured in centimeters.

Medium grained diabases were encountered in the form of dikes and necks in lower Silurian schists near Troitsky settlement. The marginal parts of these dikes consist of fine-grained diabases, which do not usually differ from the varieties of pillow lavas. Situated among the pillow lavas at the village of 3rd Yumaguzino are huge "pillows", more accurately covers, (with thickness of 1 to 2 m and extending up to 5 to 8 m) which are also formed by medium

TABLE 1. Chemical composition of a diabase from the hanging wall side of the Blyavinsky deposit and spilites connected with it.

Components	Sample no. (see notes)					
	1		2		3	
	weight percent	mol. no.	weight percent	mol. no.	weight percent	mol. no.
SiO_2	48.98	816	49.51	824	45.71	761
TiO_2	1.80	023	1.90	024	0.87	011
Al_2O_3	15.14	148	15.14	148	16.82	165
Fe_2O_2	4.12	026	2.49	016	4.14	026
FeO	7.57	106	8.14	013	6.36	089
MnO	0.23	003	0.18	003	0.19	003
MgO	6.64	164	5.62	139	5.21	129
CaO	8.36	149	6.34	112	11.88	212
Na_2O	1.88	031	5.26	085	2.37	039
K_2O	0.22	002	0.27	003	0.35	004
$\text{H}_2\text{O}+110^\circ$	3.18		2.95		3.70	
$\text{H}_2\text{O}-110^\circ$	1.82		0.98		2.36	
loss on ignition	-		1.06		-	
S	-		0.06		-	
P_2O_5	-		0.28		-	
Total	99.94		100.18			99.96

Numerical characteristics according to A.N. Zavaritsky

	Sample no. (see notes)		
	1	2	3
<i>a</i>	4.8	12.4	6.5
<i>c</i>	8.3	4.2	9.1
<i>b</i>	26.1	23.8	27.0
<i>s</i>	60.8	59.6	57.4
<i>n</i>	94	97	91
<i>f'</i>	45	44	40
<i>m'</i>	46	41	35
<i>c'</i>	9	15	25

Sample 1. Medium grained diabase with labradorite. Blyavinsky deposit, hanging wall side, drillhole 15, depth 52m. Zavaritsky's article [1]. Sample 2. Spilite with hyalo-ophitic texture. Specimen 1/1939, of the region of the Blyavinsky deposit, 2nd district. The analysis carried out in Labor. Instit. Geol. Studies, Acad. stud. U.S.S.R. [IGN AN SSSR]. Sample 3. A microlitic variety. Specimen 70/9/1934 at the electric power station in the neighborhood of the village of Herzonka. The analysis was carried out in the Labor. of Ts. N. I. G. R. I.

ained diabases. This same rock type forms this place a flow on keratophyric agglomerate which was mentioned above (fig. 2). The external appearance of the spilites here described is fairly diverse. In the fresh rocks, greenish-grey and grey tints prevail. However, in outcrops we see more often brown and yellow-brown rocks. In diabases light, white, or yellowish-white laths are discernible among the green-grey colored minerals.

Mineralogical composition: Plagioclase and monoclinic pyroxene appear to be the main constituents of these rocks. Decomposed glass is always present.

In almost all the more crystalline spilites this type plagioclase does not look very fresh. Sometimes laths of it are completely replaced by an aggregate of very small tablets of prehnite. However, in the majority of cases secondary dulled (argillized) albite is speckled with such tablets. In individual laths the relative number of tablets reaches 50 percent. Where there are fewer of them and where their dimensions are measured in several hundredths of a millimeter, it can be seen that they are oriented in the same direction.

The optical properties of albite indicate that the content of An in it does not exceed 10 to 12 percent. In albite twins with corresponding α symmetrical, $B \wedge Z = 12$ to 17° . In doubly simple Carlsbad twins, $B \wedge Z = 75$ to 76° , $B \wedge Y = 18$ to 16° , $B \wedge X = 80$ to 83° , An = 7 to 10 percent. In several fine grained diabases in all microlitic spilites the albite is fresher, i. e. there are no tablets of prehnite and it is hardly argillized.

Basic plagioclase was observed only in one specimen of a diabase, but here it has already been partially replaced by albite. The nature of the substitution is not clear, since the albite is not encountered in the same crystal with basic plagioclase. This specimen with partially argillized plagioclase was taken from the central parts of the bed in Novaya Rakityanka. In other specimens from this same bed only albite was found. The basic plagioclase is zoned. In the central and, moreover, large part of the laths appeared to be acid labradorite or basic anorthesine (40 to 50 percent An), judging from the maximum angle of extinction in the zone $[010]$, equal to 23 to 27° , and from the data of the measurement of a single simple twin $B \wedge Z = 78^\circ$, $B \wedge Y = 49^\circ$, $B \wedge X = 43^\circ$, $B = \perp [010]$ in the plane of (010) ; percent An = 48). In a very narrow peripheral zone the composition of plagioclase reaches the composition of basic oligoclase (25 percent An), judging from the angle of extinction which does not exceed 10° .

In several specimens monoclinic pyroxene appears to be the prevailing mineral. It is

found in equant, almost rounded grains, the size of which differs according to the degree of crystallinity of the rocks.

The optical properties of the pyroxene are: $\gamma = 1.718$, $\beta = 1.695$, $\alpha = 1.691$, $\gamma - \alpha = 0.025$ to 0.028 ; $2V = +48$ to 50° . $C \wedge Z$ is usually near 41° , but varies from 32 to 44° . Strong dispersion $\rho > \nu$; in one grain an inclined dispersion was observed.

Comparing these properties with data from the diagram of Winchell [162] and Tomita [163], we see that the pyroxene appears to be pigeonite common for basic extrusive rocks. The ratio of FeO:MgO = 25:75. In weight percentages FeSiO₃ = 25, CaSiO₃ = 30, MgSiO₃ = 45.

In several fine grained diabases the pyroxene possesses a distinct zoning, and sometimes a clear hourglass structure. The periphery of the small grains in these cases possesses a slightly larger angle of extinction (43 to 44° instead of 40 to 41° in the center), but $2V$ does not change. Sometimes a faint brownish tint is noticed in the periphery. As for crystallization it is evident that the composition of pyroxene moved in the direction of increasing Fe₂O₃ and TiO₂ content.

The pyroxene is usually well preserved, but sometimes it is replaced by chlorite. No indications of uraltization were observed.

Titano-magnetite is present in all of these rocks in the usual quantity for an accessory mineral. There is not much more of it in diabases, and only isolated small grains are found in microlitic rocks. In medium grained diabases the grains of titano-magnetite are xenomorphic in relation to plagioclase and pyroxene, but in other rocks they are idiomorphic.

Besides the main minerals that have been pointed out, there were encountered in the rock being described (in order of abundance): chlorites, prehnite, calcite, epidote, quartz, potash-feldspar, albite, actinolite, pyrite, hydroxides of iron. Several of them are almost always present and in a large quantity, while others were encountered only in the form of occasional grains. A large part of these are of hydrothermal origin.

The chlorites in these spilites, just like those in the rocks described above, replace mainly the glass. Therefore in several microlitic varieties the quantity of them reaches 30 to 40 percent. They are found in the form of an aggregate of very fine irregularly arranged scales. In fresh rocks the color of this aggregate is green; in partially weathered rocks, a yellowish brown.

In diabases the scales of chlorite are larger (up to 0.02 mm), but the structure of the ag-

gregate is more often concentrically zoned or spherulitic. The optical properties of the chlorite are: $\beta = 1.59 \pm$ in green varieties and up to 1.64 in yellow varieties; $\gamma - \alpha$ varies correspondingly from 0.005 to 0.015; the elongation is positive, and sometimes an inclined extinction is noted (up to 50°). Apparently, they are delessites, as was also the case in the previous spilites.

The change of the composition in the chlorites is probably connected with weathering. However, there were also observed such concentrically zoned aggregates, where the outer zones consisted of green chlorites, and the inner zones of yellow chlorites. This points to the fact that the composition of the chlorite has already shifted during crystallization.

Sometimes, instead of the ordinary delessites, a greenish-yellow mineral very similar to them is encountered, with $\beta = 1.62 \pm$, $\gamma - \alpha$ up to 0.030 and with positive elongation. It is probably a mineral of the viridite type. Another variety of chlorite is encountered in amygdulites together with calcite. It is almost colorless, with $\beta = 1.58 \pm$, $\gamma - \alpha$ less than 0.005, with a notable anomalous bluish tint in the interference color and with a positive, but sometimes also a negative elongation. This is obviously penninite. In spilites from the neighborhood of the deposit of Yaman-Gas, it also forms a matrix, in which are found only microlites of albite and small grains of secondary quartz.

Prehnite is very widespread in spilites. It is encountered in the form of the already mentioned tablets in the plagioclase; sometimes it displaces the chloritic matrix, or forms fine (up to 1 mm) veinlets in the diabases. In them tablets of the prehnite reach 0.2 mm. The optical properties of prehnite in these tablets are as follows: $\gamma = 1.652$, $\beta = 1.623$, $\alpha = 1.617$; $\gamma - \alpha = 0.033$ to 0.035; $2V = +57$ to 60° ; $\gamma \perp$ to the cleavage (001) and, consequently, the elongation of the tablets is negative.

Calcite, quartz and epidote fill out amygdulites. The calcite associates with the chlorite, and the quartz either with the epidote or with the viridite that has already been mentioned. In several rocks small grains of secondary quartz are encountered in a chloritic matrix.

Potash-feldspar was observed only in two thin sections in micropegmatitic intergrowths with quartz. The diabases with these "micropegmatites" do not resemble ordinary quartz diabases. They are undoubtedly hydrothermally changed rocks. The large quantity of quartz (about 20 to 30 percent) attracts our attention. Sometimes laths of plagioclase (albite) are located as if in one large crystal of quartz, and here they are always surrounded by a rim of micropegmatite. In the rocks there is a lot of chlorite with high birefringence. Pyroxene is absent in one thin section. Apatite, small needles of which penetrate the plagioclase, chlorite, quartz, and micropegmatite, is found

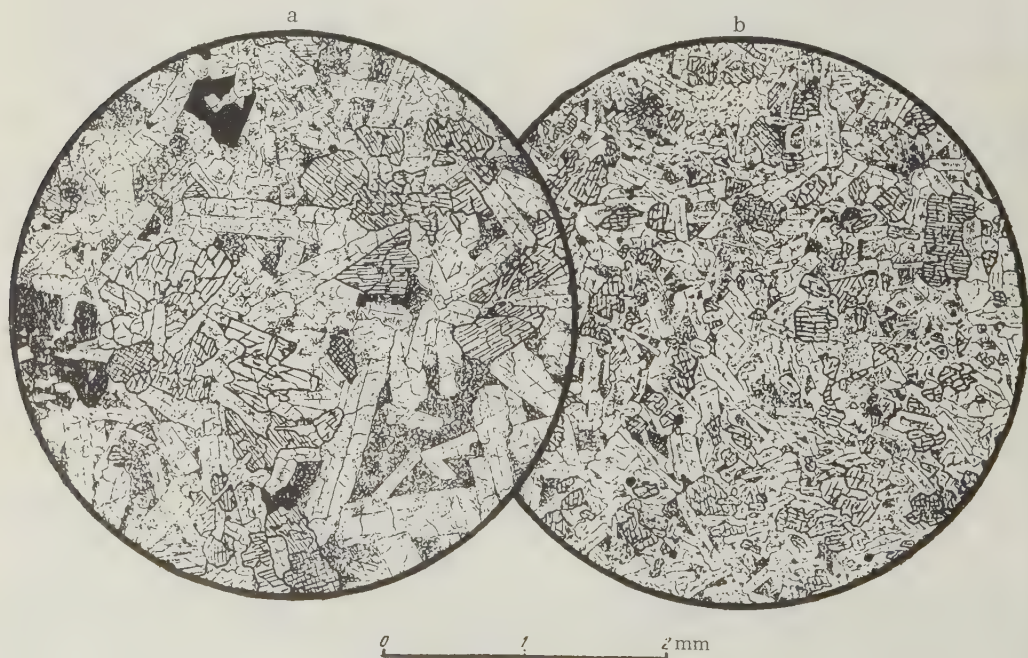


FIGURE 12. Diabase of the normal type; a) medium-grained, b) fine-grained. Laths of albite that have become turbid (covered with dots in the figure), small pyroxene grains with clear relief, and a chlorite matrix. Idiomorphism of pyroxene is more pronounced in the fine-grained variety.

re in large quantities.

Albite of hydrothermal origin is encountered in several thin sections. It is found in amygdules, and more rarely displaces the chloritic matrix or grows in the form of rims on the laths of already albitized plagioclase, adding to them irregular and crooked outlines. The albite of these rims is usually oriented perpendicularly with the albite of the laths and forms with them a single whole. However, cases were observed where the albite was oriented in the border otherwise than in the laths were also observed. The optical properties of the albite from amygdules the following were determined: $n = 1.528 \pm 0.002$ and $2V = (+) 78$ to 80° .

The texture and characteristics of the individual varieties: A character of the texture of diabases of this type appears to be the tendency of the pyroxene towards idiomorphism. Its grains are usually larger than the laths of plagioclase, despite the fact that they are found in the spaces between them. The idiomorphism of the pyroxene is more pronounced in fine grained rocks, where the rounded small grains of this mineral are sometimes arranged independently of the laths of plagioclase. In medium grained rocks the grains of pyroxene are more irregular and

sometimes separate laths of plagioclase cut into them.

The typical texture of the diabases is doleritic (fig. 12). The glass, decomposed into an aggregate of chlorite, is not found in large quantities, and sometimes cannot be found at all. Therefore a tholeiitic and intersertal texture is not common. Still less common are ophitic and poikilo-ophitic textures. However, the latter was nevertheless observed in several specimens originating from the contact of diabasic dikes.

The size of the grains of pyroxene in medium grained rocks is 0.5 to 0.8 mm, and of the laths of plagioclase from 0.3 x 0.1 to 1.0 x 0.3 mm. Plagioclase and pyroxene are encountered in almost equal quantities. In small grained diabases the small grains of pyroxene are from 0.1 to 0.2 mm in size, and the laths of plagioclase are 0.05 to 0.1 mm. There is sometimes more pyroxene in these rocks than plagioclase (up to 60 percent).

Of the microlitic varieties, the nearest to diabases is a rock in which there is almost no glass and in which pyroxene predominates over plagioclase. The small grains of pyroxene here

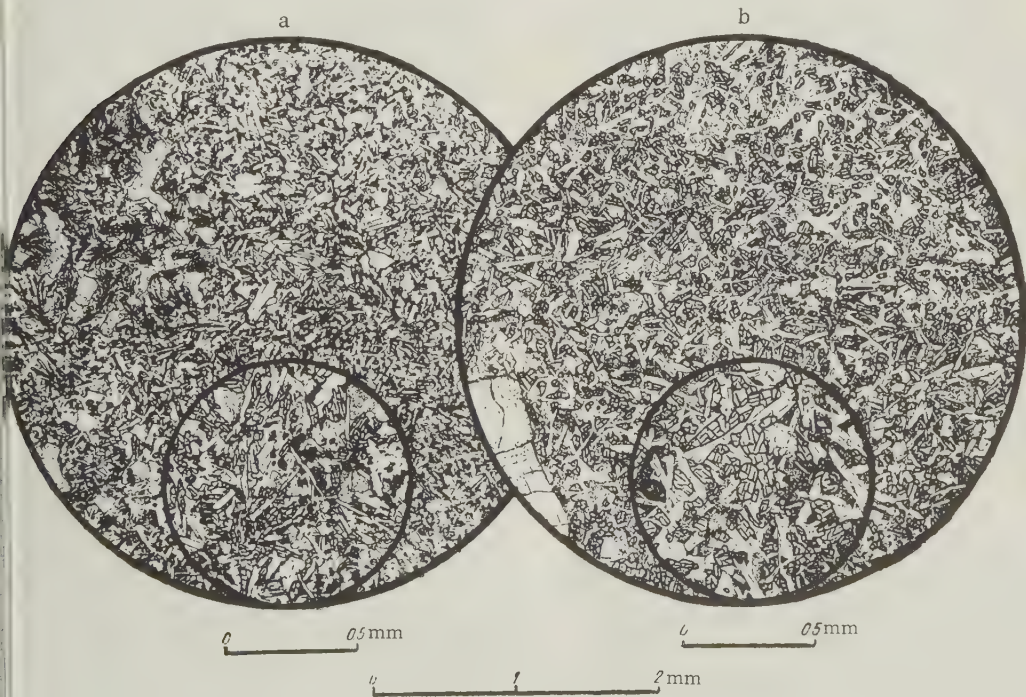


FIGURE 13. (a) A microlitic variety of a spilite from the pillow lavas in a railway cut at 232 km. In the upper part on the left the rounded portion consists of "glass" with paniced shapes. Another microlitic variety of a spilite from the neighborhood of the village of Herzonka. There are more small pyroxene grains with strong relief. On the lower left is a part of a random phenocryst of albite. The same rocks are enlarged in the small circles.

are larger than microlites of plagioclase and are always less idiomorphic than in diabases. On the whole the texture is micro-tholeiitic (fig. 13b). Individual phenocrysts of pyroxene (up to 0.5 mm) or of plagioclase (up to 2 mm) are encountered here very rarely.

The more widespread microlitic spilites, particularly in the pillow lavas in the railway cut at 232 km, differ from the rock described above by their large content of glass and the smaller dimensions of grains of pyroxene. The latter are smaller than the microlites of the albite and sometimes only the finest brown semi-transparent lumps are found instead of them. The texture of the rock is intersertal, but differs from a typical basaltic texture by the large quantity of glass (fig. 13a). Occasionally larger laths (up to 0.5 mm) of plagioclase may be taken for porphyritic phenocrysts.

One of the features of the texture of this rock appears to be the rounded formations of decomposed glass and the brown paniced bunches found in it. The latter appear to be groups of fibers of albite, the same as those which make up variolites.

The arrangement of the bunches indicates that in place of these rounded formations in semi-hardened lava there were, so to say, drops of melt, and that crystallization in these drops began from their periphery (fig. 13a). The diameter of these original sections is usually about 1 mm but sometimes it reaches 3 to 5 mm.

The least crystalline spilites of this type, encountered only in the periphery of flows or in isolated pillows of globular lavas, possess a texture, which can possibly be called hyalopilitic. The basic mass of these rocks consists of very fine microlites and a brown semi-transparent matrix. The microlites are not more than 0.08 to 0.1 mm long, and they are arranged irregularly and rather sparsely in the matrix. Occasional phenocrysts are produced by small equant grains of pyroxene (up to 0.1 mm) and by the same small laths of plagioclase.

Amygdulites are very often encountered. Around them can be seen a wide zone of a single semi-transparent matrix in which are found only small bubbles stretched in the direction of the center of the amygdulite. Sometimes the amygdulites are of a completely irregular shape, but the glassy sections around them are always rounded (fig. 14). In several rocks there are no amygdulites as such, but rounded sections are nevertheless present with radially inclined bubbles. Often the glassy matrix is strongly iron-stained and represents a semi-transparent reddish brown or completely opaque black mass.

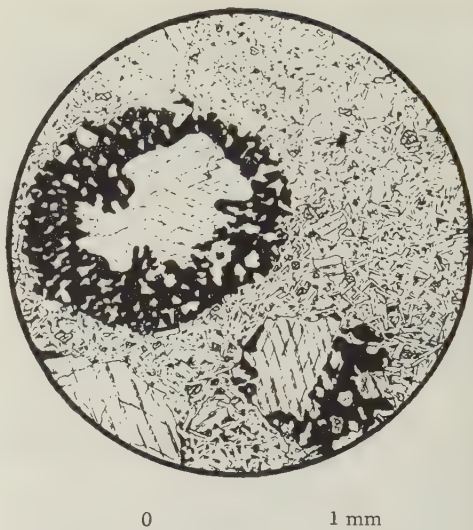


FIGURE 14. An amygdaloidal rock from the neighborhood of the village of Alchembayevo. A glassy zone enriched with magnetite is around the amygdulites. There is a radial arrangement of bubbles in it. The amygdulites are filled with calcite; there is much of it also in the main mass.

Chemical composition: Of all the varieties of spilites of the normal type that have been analysed, only the two specimens cited below are of a medium grained diabase and a microlitic spilite (table 2).

These data indicate that both varieties are related in chemical composition to rocks of a spilitic type, with a higher ratio of alkalis to feldspathic lime. However, their composition differs a little from the average composition of spilites. Our attention is drawn to the relatively high content of K_2O in the composition of the medium grained diabase.

Variolites from the neighborhood of the village of Usergan

A feature of the spilites of this type appears to be that they consist exclusively of faintly crystallized varieties, despite the comparatively large dimensions of the extrusive bodies formed by them.

In composition these rocks apparently correspond to normal spilites, but it is impossible to consider them a variety of the latter. Geologically they form an independent group, and there are also no rocks which, at least in petrographic features, have appeared to be transitional between variolites and normal spilites.

Variolites have an even distribution on the right side of the Blyava to the north and northwest of the village of Usergan and up to the northeastern outskirts of the town of Mednogorsk. Here they form a thick layer between siliceous schists. In the outcrops they are largely de-

TABLE 2. Chemical composition of spilites of the normal type.

Components	Sample no. (see notes)			
	4		5	
	weight percent	mol. no.	weight percent	mol. no.
SiO ₂	49.36	822	52.46	874
TiO ₂	1.30	016	0.94	012
Al ₂ O ₃	15.23	149	15.33	150
Fe ₂ O ₃	3.56	023	4.09	026
FeO	6.62	092	6.13	085
MnO	0.20	003	0.18	003
MgO	6.76	168	4.69	117
CaO	8.00	143	5.98	107
Na ₂ O	3.32	053	5.62	090
K ₂ O	1.74	018	0.19	002
H ₂ O+110°	2.57		3.81	
H ₂ O-110°	0.58		0.81	
P ₂ O ₅	0.23		-	
Loss on ignition	0.53		-	
S	0.08		-	
BaO	0.10		-	
Total	100.18		100.23	

Numerical characteristics according to A.N. Zavaritsky.

	Sample no. (see notes)	
	4	5
<i>a</i>	9.9	12.8
<i>c</i>	5.4	4.0
<i>b</i>	26.1	21.3
<i>s</i>	58.6	61.9
<i>n</i>	75	98
<i>f'</i>	38	46
<i>m'</i>	45	38
<i>c'</i>	17	16

Sample 4. Medium-grained diabase from the thick cover at the village 3rd Yumaguzino, specimen 9/1939. The analyses were carried out in the laboratory of IN. AN SSSR (Petroleum Institute of the Academy of Sciences of the U.S.S.R. Ed.) by V. A. Egorova. Sample 5. Microlitic spilite of yellow lava from the railway cut 232 km., specimen 37. From an article by Zavaritsky [1].

destroyed, and it is difficult to define the forms of the extrusive bodies. There are no typical yellow lavas here, but on several slopes of hills huge flat slightly isolated pillows can be seen, with a thickness of 1.5 to 3 m, and a long axis of 8 to 10 m. The center and periphery of these bodies are composed of one rock type, although in the periphery, glassy varieties are observed.

Besides in the neighborhood of the village of Ussergan, spilites of this type are encountered on the right bank of the Kuragan River, 1 km below the estuary of the Blyava.

Macroscopic variolites have the appearance of an aphanitic dense rock, most often grey in color with a greenish tint and reddish-brown irregular spots. Sometimes there are so many of these spots that they merge, and the rock takes on a reddish-brown coloring. This mottledness is caused by an uneven distribution of oxides of iron and is not dependent upon the micro-texture of the rock.

In weathered variolites the coloring is a dirty, yellowish-grey. At first glance these rocks resemble keratophyres.

Microscopic features of variolites: Under a microscope it is possible to distinguish in almost all variolites of the type being considered, a main mass, which possesses a radial-fibrous structure, and numerous, fine idiomorphic, small grains (evenly dispersed in it) of

monoclinic pyroxene (fig. 15). They are distributed completely independently of the arrangement of the plagioclase fibers in the main mass, taking no part in the structure of the separate bundles. The small grains of pyroxene have the shape of prismatic crystals, in which the common faces (010) are absent. The cleavage cannot be recognized even with the largest magnifications. In the main body of the small prisms there is almost always found a yellow-brown aggregate of chlorite, consisting of glass caught during crystallization and subsequently decomposed. The overall length of the small prisms is 0.1 to 0.2 mm and sometimes reaches 0.5 to 0.8 mm. They are about 0.03 to 0.05 mm in diameter.

The optical properties of pyroxene are determined with great difficulty because of the small dimensions of the grains. The pyroxene is completely colorless, $\beta = 1.693 \pm$, $2V = +44$ to $+50^\circ$, $C \wedge Z = 38$ to 44° . In the largest prisms a zoning and hour-glass structure were observed. In the main body, $2V$ is larger than in the periphery, while $C \wedge Z$, on the contrary, is less. Judging from these data the composition of the pyroxene in the variolites is apparently the same as that in normal spilites, i. e. it corresponds to the composition of pigeonite. During alteration, the pyroxene changes into green chlorite, with the transformation starting from the ends of the small prisms. The character of the distribution of the pyroxene and its distinctive shapes give the variolites being described a similarity with



FIGURE 15. Variolites from the neighborhood of the village of Usergan. a) Atypical texture of the b) The same greatly magnified. The array of the small pyroxene grains can be seen, which is independent of the array of albite fibers. A chloritic aggregate ("glass") is in the center of the small grains.

the world famous rocks of Yalguba.

The principal feature of the main mass appears to be the absence of isolated varioles and an intervariolithic mass. It consists wholly of bunches of fibers of plagioclase. The thickness of the fibers is about 0.002 to 0.005 mm, so they are distinguishable only under very high magnification. Their length varies from 0.2 to 0.3 mm to 1.0 to 1.5 mm. Sometimes they are bent. In an isolated bunch, submicroscopic particles of the products of decomposition are tangled between the fibers, thanks to which the fibrous structure is noticeable even without an analyser.

The plagioclase in the fibers always consists of albite since the index of refraction is less than that of Canada balsam.

The most common shape of the bunches is panicked. It differs from a radial or spherulitic shape in as much as the fibers do not go out from a single point, but from a certain line. Besides the panicked bunches, there are encountered sheaf-like and parallel groups of fibers. In rocks with a typical texture (fig. 15a), the arrangement of the bunches is irregular. Often they are entangled and even grow through each other. Such rocks may possibly be called variolithic aphanites, the more so since they are

almost identical with the same rocks from Yalguba, to which rocks this name was at first given.

Some of the variolites being described have a texture that differs from a typical texture in that the bunches of diverse fibers are quite sparsely distributed in the mass, which looks perfectly uniform without an analyser, and which, with crossed nicols, displays a submicroscopically fine fibrous composition. The usual bunches change gradually into this mass. Besides that, they are distributed more regularly in it. Most often several panicked bunches come together approximately to a single center, often forming isolated varioles of an irregular shape.

The texture completely resembles that which Levinson-Lessing, called "radiolitic" in contrast to spherulitic. Of the variolites of Yalguba, the types that were first known there, resemble these rocks. Sometimes, in variolites with a typical texture there are encountered rounded sections in which small prisms of pyroxene are completely absent, and in which the bunches of fibers have more fantastic shapes.

The least crystalline varieties of variolites are represented by two rock types. One of them composes the very periphery of a lava bed. It is a darker glassy rock, which ought

be attributed to palaeotachylites. Under a microscope a uniform brown semi-transparent mass can be seen, in which, from time to time, are trapped pseudomorphs of chlorite in small prisms of plagioclase (0.4 to 0.2 mm) and sheaf-like bunches of coarse fibers of albite. The glassy mass is not quite isotropic. With crossed nicols there are in it faintly visible birefringent fine irregular spots, probably spherulitic formations of submicroscopically fine fibers.

Another slightly crystalline variety, the bedding conditions of which are obscure, represents a light grey aphanitic species. On a partly polished surface completely round, lighter spots are noticed with a diameter of up to 5 mm. Under a microscope these spots prove to be varioles of very fine fibers. The varioles have a typical spherulitic structure; i.e., the fibers in them come together towards a single center. The intervariolitic mass does not have a fibrous structure, but as a result of collective crystallization it is slightly anisotropic. Its darker coloring is caused by the numerous small crystals of magnetite.

The main feature of the rock appears to be the absence in it of small prisms of pyroxene. Instead, there were observed occasional pseudomorphs of calcite in sizes up to 0.2 mm. Their outlines correspond to the outlines of the small prisms of pyroxene. Decomposed albite is also found in their centers.

Chemical composition: The chemical analysis of the variolite from the neighborhood of the village of Usergan is given in Table 3.

TABLE 3. Chemical composition of a variolite from the neighborhood of the village of Usergan

Components	Sample no. (see notes) 6	
	weight percent	mol. no.
SiO ₂	57.70	961
TiO ₂	0.29	004
Al ₂ O ₃	16.01	157
Fe ₂ O ₃	2.49	016
FeO	3.90	054
MnO	0.08	001
MgO	5.65	140
CaO	6.85	122
Na ₂ O	4.44	072
K ₂ O	0.47	005
H ₂ O+110°	3.26	
H ₂ O-110°	0.66	
Total	101.80	

Numerical characteristics according to A.N. Zavaritsky.

Sample no. (see notes) 6	
<i>a</i>	= 10.5
<i>c</i>	= 5.5
<i>b</i>	= 18.3
<i>s</i>	= 65.7
<i>n</i>	= 94
<i>f'</i>	= 32
<i>m'</i>	= 52
<i>c'</i>	= 16

Sample 6. Variolite specimen 18 [14] 1934, from the neighborhood of the village of Usergan. The analysis was carried out in the laboratories of Ts. N. I. G. R. I., (Control Scientific Research Institute of Geological Survey, Ed.).

These data indicate that the variolites here described are nearer in chemical composition to rocks of an andesitic-basaltic type than to

spilites. The ratio of feldspathic lime to alkalis in them is larger despite the fact that the plagioclase is also albite.

Variolites from the neighborhood of the villages of Herzonka and Alchembayevo

There are not many data about these spilites. The geological conditions of their bedding are not clear, although it is known that to the south of the village of Herzonka and up to the village of Alchembayevo they form a thick layer which lies on siliceous schists. The transitions of these rocks into spilites of some other type were not observed. Geologically they are also an independent group, although in composition these variolites are close to the spilites from the neighborhood of the Blyavinsky deposit.

The external appearance of the rocks does not attract attention. They are strongly fissured, brownish-grey aphanitic rocks with numerous veins of calcite. Under a microscope, however, it strikes us at once that the pyroxene here takes part in the structure of the bunches of fibers. It is found in the form of very fine (0.01 to 0.03 mm) grains or needles arranged only between the fibers of plagioclase. The latter here are coarser than the foregoing type, and, being separate small grains of pyroxene, they are quite different from the other.

The plagioclase in the fibers appears somewhat dimmed by the albite. In a single thin section it consists of labradorite, since here its index of refraction is greater than in

Canada balsam.

Besides the fibers of plagioclase and the



FIGURE 16. A variolite from the neighborhood of the village of Herzonka. a) The typical texture of the rock. b) The same, greatly magnified. One can see the dependence of the arrangement of small pyroxene grains on the arrangement of albite fibers.

small grains of pyroxene in the structure of the bunches there is also present a greenish-yellow chlorite representing a product of decomposed glass. The shape of the bunches is mainly panicled although spherulitic groups are encountered more often in the spilites being described than in the preceding ones. The fibers here are rectilinear, 1.5 to 2.0 mm in length. Where the two bunches touch the contact line between them has the appearance now of a straight line and now of a broken line.

The absence of varioles and of an inter-variolic mass is still on the whole characteristic for the texture of rocks here described. They are still more uniform than the previous variolites, since they are formed wholly from identical bunches (fig. 16).

Of the other features there ought to be noted the pseudomorphs of serpentine and calcite, which by their outlines bring to mind crystals of olivine with faces (021). Sometimes a mesh-texture can be seen in them. Their dimensions are from 0.2 to 0.5 mm. They are often found in the center of a spherulitic bunch.

A somewhat unusual texture was observed in a specimen of a variolite from the exposed spilitic agglomerates in the railway cut at

the village of Alchembayevo. Here strong curvature of the fibers is characteristic (fig. 17), as is also the absence of small grains of pyroxene. Between the fibers there are now submicroscopic particles.

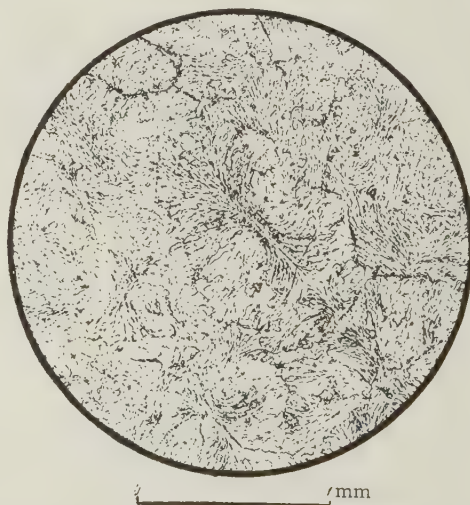


FIGURE 17. Variolitic aphanite from the spilitic agglomerates in the environs of the village of Alchembayevo. Pyroxene is absent. The strange shape is that of panicles.

ASYMMETRIC INTERFERENCE WAVES IN A LAMINATED MEDIUM¹

by V. I. Keylis-Borok

• translated by H. P. Thielman •

ABSTRACT

Investigation of axial-symmetric interference waves in a multilayered half space is extended to the problem of asymmetric oscillations. The solution presented is in analogous form to that of the author's axial-symmetric problem. The method of solution is based on a problem that involved a homogeneous half space: the solution is expressed as Fourier double integrals in terms of x , y ; and, then, reduced to single integrals of the Fourier-Bessel type. In the asymmetric case of excitation, analysis of formulas expressing displacement shows the principal component of the displacement to be separated into waves of two types: 1) Those analogous to axial-symmetric waves, whose displacements lie in the vertical plane; and 2) those having horizontal, tangential displacements, a special case of Love waves. Apparently, simultaneous generation of these wave types is investigated for the first time in this work. Their similarity to volume waves is substantiated here. The relations derived make it possible to separate the influence of properties of the medium from those of the source upon the asymmetry of surface waves (in particular, upon the relationship between Rayleigh and Love waves). Possibly, this study might be applicable to research on surface-wave absorption in the earth's crust as well as on dynamical properties of a disturbance focus. The formula given for Love-wave intensity is considered to be especially applicable to investigation of the structure of laminated media, on the basis of resonance phenomena. --D. D. Fisher.

In earlier works [1, 2] the author has investigated axial-symmetric interference waves in a multi-layered half-space.

In the present work, an attempt is made to extend the results obtained earlier to a symmetric oscillations. The statement of the present problem is identical to that given in the paper [2] except for the condition of symmetry.

We shall make use of the same notation as that used in the earlier works [1, 2] and introduce the following additional symbols: $w_q^{(1)}$, $w_q^{(2)}$, $w_q^{(3)}$ are the components of displacement in the q -th layer along the axes x , y , z , respectively; $F_q^{(1)}$, $F_q^{(2)}$, $F_q^{(3)}$ the components of the stresses applied at the q -th boundary; $Q_q^{(j)}$ is the spectral function in the representation of $F_q^{(j)}$ as a double Fourier integral.

In order to obtain a solution in analogous form to solution of the axial-symmetric problem, we make use of a method developed in an example involving a homogeneous half-space in the work of reference [2]: we express the solution as Fourier double integrals in terms of x , y , and then reduce them to single integrals of the Fourier-Bessel type.

By separating the variables one can write $w_q^{(j)}$ as:

$$w_q^{(j)} = \frac{1}{2\pi i} \int_{-\infty}^{\infty} U(p) e^{ipt} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \frac{f_q^{(j)}(m, n, p, z)}{\Sigma_n(m, n, p, z)} e^{i(mx+ny)} dm dn dp. \quad (1)$$

Here

$$\Sigma_n = \prod_{q=1}^n \alpha_q \beta_q^2 |K_0 K_1 \dots K_n| \quad (2)$$

¹Translated from *Asimmetrichnyye interferentsionnyye volny v sloistoy srede*; Doklady Akademii Nauk SSSR, Novaya Seriya, v. 107, no. 4, 1956, p. 533-536; presented by Academician O. Yu. Schmidt on 27 July 1955.

is the determinant of the system of equations for the boundary condition; the K_j are matrices given by the formulas (3) to (5) where numbers in parentheses indicate the numbers of rows and columns; the omitted elements are all zero:

$$\xi^2 = m^2 + v^2; \gamma_j = 2\xi^2 - k_j^2; \sigma_0 = -1;$$

$$l = \text{ch} \alpha_{j+1} z_{j+1}; u = \alpha_{j+1} \text{sh} \alpha_{j+1} z_{j+1}; v = \frac{\text{sh} \alpha_{j+1} z_{j+1}}{\alpha_{j+1}};$$

the dash (ˉ) above l , u , v indicates that z_{j+1} is to be replaced by z_{j+2} ; the prime (') on l , u , v indicates that α is to be replaced by β . The remaining symbols are the same as those given in the earlier work [2].

$$K_0 = \begin{vmatrix} \begin{matrix} (1) & (2) & (3) & (4) & (5) & (6) \\ -2i\xi^2 u & -2i\xi^2 l & \gamma_1 v l & -\gamma_1 v v & -\gamma_1 m l & \gamma_1 m v \\ 0 & 0 & \beta_1^2 m l & -m u & \beta_1^2 v l & -v u \\ -\gamma_1 l & \gamma_1 v & 2i v u & -2i v l & 2i v u & -2i m l \\ -2i\xi^2 \bar{u} & 2i\xi^2 \bar{l} & \gamma_1 v \bar{l} & -\gamma_1 v \bar{v} & -\gamma_1 m \bar{l} & \gamma_1 m \bar{v} \\ 0 & 0 & \beta_1^2 m \bar{l} & -m \bar{u} & \beta_1^2 v \bar{l} & -v \bar{u} \\ -\gamma_1 \bar{l} & \gamma_1 \bar{v} & -2i v \bar{u} & 2i v \bar{l} & 2i m \bar{u} & -2i m \bar{l} \\ i\xi^2 \bar{l} & -i\xi^2 v & -v \bar{u} & v \bar{l} & m \bar{u} & -m \bar{l} \\ 0 & 0 & -m \beta_1 \bar{l} & m \beta_1 \bar{l} & -v \beta_1 \bar{l} & v \bar{l} \\ -\bar{u} & \bar{l} & -i v l & i v \bar{v} & i v \bar{v} & -i m \bar{v} \end{matrix} \end{vmatrix} \quad (3)$$

$$K_j = \begin{vmatrix} \begin{matrix} (6j+1) & (6j+2) & (6j+3) & (6j+4) & (6j+5) & (6j+6) \\ 2\sigma_j i \xi^2 u & -2\sigma_j i \xi^2 l & \sigma_j v l' & \sigma_j v v' & \sigma_j \gamma_{j+1} m l & -\sigma_j \gamma_{j+1} m v' \\ 0 & 0 & -\sigma_j \beta_{j+1}^2 m l' & \sigma_j m u' & -\sigma_j \beta_{j+1}^2 v l' & \sigma_j v u' \\ \sigma_j \gamma_{j+1} l & -\sigma_j \gamma_{j+1} v & 2\sigma_j i v u' & -2\sigma_j i v l' & -2\sigma_j m u' & 2\sigma_j i v l' \\ -i \xi^2 l & i \xi^2 v & v u' & -v l' & -m u' & m l' \\ 0 & 0 & m u' & -m l' & v u' & -v l' \\ u_i & -l_j & i v l' & -i v v' & -i m l' & i m v' \\ -2i \xi^2 \bar{u} & 2i \xi^2 \bar{l} & \gamma_{j+1} v \bar{l}' & -\gamma_{j+1} \bar{u}' & -\gamma_{j+1} m \bar{l}' & \gamma_{j+1} m \bar{v}' \\ 0 & 0 & \beta_{j+1}^2 m \bar{l}' & -m \bar{u}' & \beta_{j+1}^2 v \bar{l}' & -v \bar{u}' \\ -\gamma_{j+1} \bar{l} & \gamma_{j+1} \bar{v} & -2i v \bar{u}' & 2i v \bar{l}' & 2i m \bar{u}' & -2i m \bar{l}' \\ i \xi^2 \bar{l} & -i \xi^2 \bar{v} & -v \bar{u}' & v \bar{l}' & m \bar{u}' & -m \bar{l}' \\ 0 & 0 & -m \bar{u}' & m \bar{l}' & -v \bar{u}' & v \bar{l}' \\ -\bar{u} & \bar{l} & i v \bar{l}' & i v \bar{v}' & i m \bar{l}' & -i m \bar{v}' \\ \dots & \dots & \dots & \dots & \dots & \dots \end{matrix} \end{vmatrix} \quad \begin{matrix} (6j-2) \\ (6j-1) \\ (6j) \\ (6j+1) \\ (6j+2) \\ (6j+3) \\ (6j+4) \\ (6j+5) \\ (6j+6) \\ (6j+7) \\ (6j+8) \\ (6j+9) \end{matrix} \quad (4)$$

$$j = 1, \dots, n-1;$$

$$K_n = \begin{vmatrix} 2\sigma_n i \xi^2 \alpha_{n+1} - \sigma_n \gamma_{n+1} \nu & \sigma_n \gamma_{n+1} m \\ 0 & -\sigma_n \beta_{n+1}^2 m - \sigma_n \beta_{n+1}^2 \nu \\ \sigma_n \gamma_{n+1} & -2i\sigma_n \beta_{n+1} \nu & 2i\sigma_n \beta_{n+1} m \\ -i\xi' & \beta_{n+1} \nu & -\beta_{n+1} m \\ 0 & \beta_{n+1} m & \beta_{n+1} \nu \\ \alpha_{n+1} & i\nu & -im \end{vmatrix}; \quad (5)$$

$$f_q^{(1)} = im\Phi_q - \beta_q \bar{\Psi}_q^{(2)}; \quad f_q = i\nu\Phi_q + \beta_q \bar{\Psi}_q^{(1)}; \quad f_q^{(3)} = \alpha_q \bar{\Phi}_q - i\nu\Psi_q^{(1)} + im\Psi_q^{(2)};$$

$$\begin{aligned} \bar{\Phi}_q &= A_{6q+1} \operatorname{ch} \alpha_q z + A_{6q+2} \frac{\operatorname{sh} \alpha_q z}{\alpha_q}; \\ \Phi_q &= A_{6q+1} \operatorname{sh} \alpha_q z + A_{6q+2} \frac{\operatorname{ch} \alpha_q z}{\alpha_q}; \\ \Psi_q &= A_{6q+2+i} \operatorname{ch} \beta_q z + A_{6q+3+i} \frac{\operatorname{sh} \beta_q z}{\beta_q}; \\ \bar{\Psi}_q^{(i)} &= A_{6q+2+i} \operatorname{sh} \beta_q z + A_{6q+3+i} \frac{\operatorname{ch} \beta_q z}{\beta_q}; \\ \Phi_n &= \bar{\Phi}_n = A_{6n+1} e^{-\alpha_{n+1} z}; \quad \Psi_n^{(i)} = \bar{\Psi}_n^{(i)} = A_{6n+1+i} e^{\beta_{n+1} z}; \\ q &= 0, 1, \dots, n-1. \end{aligned} \quad (6)$$

A_k is the determinant obtained from Σ_n by replacing elements of the k -th column by the following quantities: zero for the $(6k+1)$ -st, $(6k+2)$ -nd, $(6k+3)$ -rd rows where $k = 1, 2, \dots, n$; $-\frac{1}{\mu_k} Q_k^{(j)}(\xi)$ for the rows numbered i and $(6k-1-i)$.

The study shows that:

1) Σ_n can be represented as:

$$\Sigma_n = \Delta_n(\xi) L_n(\xi), \quad (7)$$

where L_n , the determinant of the system of boundary conditions for waves of the Love type, is given by the formula

$$L_n = \xi^{3n+2} \beta_{n+1} \prod_{q=1}^n \beta_q |T_0 T_1 \dots T_n|; \quad (8)$$

$$T_0 = \begin{vmatrix} \text{(1)} & \text{(2)} \\ \beta_1^2 l & -u \\ \beta_1^2 \bar{l} & -\bar{u} \\ -u & -\bar{l} \end{vmatrix} \begin{matrix} (1) \\ (2) \\ (3) \end{matrix} \quad T_q = \begin{vmatrix} (2q+1) & (2q+2) \\ -\sigma_q \beta_{q+1}^2 l & \sigma_q u \\ u & l \\ \beta_{q+1}^2 \bar{l} & -\bar{u} \\ -\bar{u} & \bar{l} \end{vmatrix} \begin{matrix} (2q) \\ (2q+1) \\ (2q+2) \\ (2q+3) \end{matrix}; \quad T_n = \begin{vmatrix} (2n+1) \\ \sigma_n \beta_{n+1}^2 \\ \beta_{n+1} \end{vmatrix} \begin{matrix} (2n) \\ (2n+1) \end{matrix}$$

2) $f_q^{(i)}$ can be expressed as a finite sum of terms that are multiples of Δ_n or L_n .

In order to make more explicit and more understandable further results, let us assume that $Q_q^{(i)}$ depends only on ξ (i. e., that stresses on the boundary are symmetric in modulus with respect to the z -axis). Then the vertical stresses $F_q^{(3)}$ will produce axial-symmetric oscillations, described in the earlier work [1].

Without loss of generality, one can assume that the point under consideration lies on the plane $y = 0$, and, that the horizontal stresses $F_q^{(N)} = \sqrt{[F_q^{(1)}]^2 + [F_q^{(2)}]^2}$ form an arbitrary constant angle δ with the x -axis. Then the displacements caused by $F_q^{(N)}$, can be expressed as:

$$\begin{aligned} w_q^{(1)} &= \cos \delta \tilde{V}_q^{(1)} + \frac{\cos \delta}{r} \omega_q^{(1)}, & w_q^{(2)} &= \frac{\sin \delta}{r} \tilde{V}_q^{(2)} + \sin \omega_q^{(2)}, \\ w_q^{(3)} &= \cos \delta \tilde{V}_q^{(3)} + \frac{\cos \delta}{r} \omega_q^{(3)}. \end{aligned} \quad (9)$$

Here the first and second terms represent displacements produced by $F_q^{(1)}$, and $F_q^{(2)}$, respectively.

In the notation of the earlier paper [1], $\tilde{V}_q^{(1)}$, $\tilde{V}_q^{(2)}$ are identical with the horizontal $V_q^{(1)}$; while $\tilde{V}_q^{(3)}$ is identical with the vertical $V_q^{(2)}$ displacements from the source with the spectrum $Q_q^{(N)}(\xi) U(p)$.

The formula for $\tilde{V}_q^{(1)}$ coincides with $V_q^{(1)}$; while the formula for $\tilde{V}_q^{(2)}$ can be obtained from $V_q^{(1)}$, by replacing $J_1(\xi r)$ by the expression $\left[\xi J_0(\xi, r) - \frac{1}{r} J_1(\xi r) \right]$ in the integral with respect to ξ . The formula for $\tilde{V}_q^{(3)}$ is obtained from $V_q^{(2)}$ by replacing $J_0(\xi r)$ by $\xi J(\xi r)$ in the integral with respect to ξ .

The displacement $\omega_q^{(j)}$ is given by:

$$\omega_q^{(j)} = \int_{-\infty}^{+\infty} U(p) e^{ipt} \int_{e(p)} \frac{\xi M_q^{(j)}(\xi, p)}{L_n(\xi, p)} k^{(j)}(\xi, r) d\xi dp, \quad (10)$$

where $k^{(1)} = J_1(\xi r)$; $k^{(2)} = \xi J_0(\xi r) - \frac{1}{r} J_1(\xi r)$; $k^{(3)} = 2J_2(\xi r)$; $M_q^{(1)} = M_q^{(2)} = P_{2q-1}(\xi) \beta_q \text{ch} \beta_q z - P_{2q}(\xi) \text{ch} \beta_q z$; $M_q^{(3)} = P_q(\xi) \text{ch} \beta_q z - P_{2q}(\xi) \frac{\text{sh} \beta_q z}{\beta_q}$; P_i is the determinant

obtained from (8) by replacing elements of the i -th column by the following quantities: $-\frac{1}{\mu_j} Q_j^{(N)}$ in the first and all even numbered rows, and by zero in all remaining rows.

The principal part \tilde{V}_q^j of the estimate for large r coincides with the principal part of the axial-symmetric displacement given in [2]; the values $\omega_q^{(j)}$ are similar in structure but much simpler. Analysis of the formulas (9) shows that, in the asymmetric case of excitation, the principal part of the displacement breaks up into waves of two types (these waves consist of unequal numbers of harmonics that differ in dispersion and resonance properties). Waves of the first type are analogous to the axial-symmetric waves described in earlier works [1, 2]; displacements for these waves lie in the vertical radial plane. Waves of the second type have horizontal, tangential displacements; Love waves are a special case of these waves. Simultaneous generation of waves of both types by a given source is investigated here, seemingly, for the first time. The intensity of oscillations of the first type at any given point is proportional to the stress lying in the same plane as the displacement. Oscillations of the second type are proportional to the projection of stresses on the direction of displacement.

The analogy with volume waves is strengthened also by the fact that displacements in the first waves have a tangential component; while displacements in the second waves have radial and vertical components whose attenuation (damping) is much greater than that of the principal oscillations.

The relations obtained make it possible to separate the influence of properties of the medium and source upon the asymmetry observed for surface waves (in particular, upon the relationship between Rayleigh and Love waves).

The main direction along which the results obtained could be applied, is the study of surface-wave absorption in the earth's crust (separation of the disturbed zone) and, the study of dynamical properties (energy and direction) of a disturbance focus. The relative simplicity of the formula given that expresses the intensity of the Love wave, should receive special attention. It is especially well suited for use in study of the structure of laminated media on the basis of resonance phenomena.

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, Doklady Akademiy

THE CONCEPT OF "FACIES" (PART 3 OF 3)¹

by V. P. Markevich

• translated by Ivan Mittin •

CHAPTER IV. DEFINITION OF THE TERM "FACIES" AND THE TERMS RELATED TO THIS CONCEPT

Taking into consideration the basic meaning of the term facies and the contents attached to it by the majority of researchers, we propose the following definition: "A 'facies' or 'a geological facies' is some volume of a deposit or a rock which is characterized by a similar complex of paleontologic, petrographic, and physico-chemical features produced by tectonic physico-chemical, biotic, and geographic conditions of sedimentation".³

Such a definition gives the term facies quite a definite meaning and narrows its field of application. The term facies becomes definite, concrete, and clear. The tectonic factor, which plays a highly important part in facies correlation, is also taken into account here.

Attempts to use the geotectonic principle for segregation of facies have been noted in the literature since the end of the 19th century. Subsequently, both in the U. S. S. R. and abroad, this question has received much attention. Expressions such as "platform facies", "geosynclinal facies", "foredeep facies", etc., can frequently be found in the literature, which emphasizes the singularity of sedimentary processes for different geotectonic zones.

The above definition eliminates any ambiguity in the interpretation of the term facies, makes its application to the study of specific material easier, and permits segregated facies not only to be described and studied but also to be represented graphically.

For greater clarity, definitions of some features related to the concept "facies" are given below.

"Facies features" - primary features of a deposit; they are conditioned by specific features of the sedimentary processes and reflect the peculiarities of the tectonic, physico-chemical, biotic, and geographic sedimentary environment.

"Facies changes" or "facies variability" of rocks or deposits - the change of facies features in a rock or deposit extended through time and space.

"Facies conditions" - conditions of sedimentation; environmental conditions at time of deposition.

"Sedimentation process" a combination of processes related to the formation of a deposit from the moment of the separation of sedimentary material from the parent rock until its deposition.

"Sedimentation basin" - a downwarp of the earth's crust related to the formation of sedimentary strata. Such a downwarp is characterized by a distinctive complex of inherent physico-geographic, physico-chemical, and biotic conditions of distribution and formation of deposits.

"Facies complex" - a group of facies unified by certain common facies features.

"A suite" - a sedimentary rock unit comprising a group of facies or facies complexes which are characterized by common distinct facies features setting this unit off amongst other deposits within the entire sedimentation basin or a considerable part of it.

"A rock" - any natural accumulation of hard mineral matter and organic remains in the earth's crust.

"Sedimentary rock" - any natural accumulation of fragments of hard mineral matter and organic remains in the earth's crust including accumulations of chemically precipitated material from water.

"Magmatic rock" - a natural accumulation of hard mineral matter in the earth's crust which formed at the expense of the cooling of magma.

"A deposit" - a sedimentary rock in the stage of deposition under conditions of the depositional environment.

If factual material is available in sufficient quantity, then every facies--or facies-stratigraphic analysis-- must be completed by compiling maps and diagrams showing distribution of facies and strata, as well as facies sections.

"A map or diagram of facies" - a drawing

¹Translated from Ponyatie "Fatsiya": Akademii Nauk SSSR, Moskva, 1957. Edited by E. D. McKee and Curt Teichert, U. S. Geological Survey. Published with the permission of the Director, U. S. Geological Survey. Part 1 of this work was published in International Geology Review, v. 2, no. 5, p. 367; Part 2 was published in v. 2, no. 6, p. 498.

²U. S. Geological Survey.

³Such a definition was first published by this author in "Geologicheskoye stroeniye Vostochnoi Gruzii", Academy of Sciences, U.S.S.R., 1954.

showing the distribution of facies or facies complexes on a plane.

"Facies sections" - a drawing showing sections of facies or facies complexes.

Depending on the accuracy of factual material, each compiled map or diagram gives a picture of the areal distribution of a facies at a definite moment during the deposition of sediments in a given region. One can establish the pattern in the change of facies and, consequently, the conditions of sedimentation more extensively and accurately if there are more maps and diagrams of facies compiled for different moments of deposition in the region. In other words, each single map of facies is essentially a framework and its combination with others forms a system in which the facies become dynamic, alive, relative, and reflect vividly the changes of facies in time and space--the changes that were caused by changes in processes of sedimentation and conditions of these processes.

A study of all facies features in a given volume of rock or deposit is not required for the segregation of a facies. In order to solve one or another kind of problem, one must know how to separate the most essential and characteristic features of the entire aggregate of facies, and to recognize those that are typical of a given definite facies and by which a given facies is sufficiently distinguished amidst the surrounding deposits or differs from the adjacent facies. A facies is not a rigidly established volume of a deposit or rock. Naturally, in each specific case, the determination of the facies volume depends on different facies features or their complexes which are accepted as basic correlative features in segregation of facies. One ought to bear in mind that the varied activities of different factors in different sections of each depositional basin depend on special features and are reflected in the deposits. It is well known that most sediments of various types--chemical, organic, terrigenous, volcanic, etc., can form in a single depositional basin. In other words, in some sections of this basin a considerable or even decisive role in forming a deposit belongs to physico-chemical factors and the deposits formed are chemical; in the others at that time the role belongs to the biological factor, and the various layers formed are biogenic, which gradually grade into terrigenous deposits which are then characterized by different conditions of sediment accumulation, with these conversions often taking place gradually.

In the literature there are also used some definite terms, such as: "petroliferous facies", "petroleum facies", "petroleum-producing facies", "petroleum-containing facies", etc. These terms have been used in designating various conjectural phenomena in the origin of petroleum or in strata of sedimentary

rocks with which these phenomena may be or were connected in the past.

First of all, petroleum and petroleum gas are not a direct result of sedimentation and, therefore, they reflect these processes only indirectly. Petroleum and petroleum gas are a result of complex physico-chemical processes which take place in a sediment or rocks associated with the transformation of the primary accumulated organic material into petroleum and gas. Therefore, one cannot accept the term "petroliferous facies" as a combination of conditions both of sedimentation and of transformation of the source material into petroleum and gas. Thus, V. V. Weber (1947) writes: "Favorable conditions for forming the petroliferous facies depend not entirely on the proper hydrobiological and physico-geographic environment of the basin which assures the accumulation in them of the organic source material but also on subsequent geologic factors that promote the transformation of this matter into petroleum."

Still further in the same article, V. V. Weber continues: "The development of petroliferous facies both during stages of the accumulation of organic material and during its subsequent transformation follows general patterns, the entire combination of which can be brought about only with a definite sequences of facies development of a given region"(p. 5). Such an interpretation of the facies makes this concept indefinite and vague.

I. O. Brod (1951) assumes that "the majority of the petroleum parent deposits are in strata of pelite and carbonate sediments containing a considerable amount of the organic matter which has accumulated under circumstances of subsidence of the basin bottom predominant over upward motions in the process of oscillation." In his opinion, "a sedimentary environment of these rocks favorable to bitumen generation can be regarded as a source facies of petroleum. A thick rock mass originated under circumstances favorable for the formation of bitumen, corresponds to the source formation" (p. 74).

It is obvious from the above quotation that I. O. Brod interprets the petroleum source facies as "conditions of the depositional environment propitious for the formation of bitumen." However, somewhat further on the same page he says that "petroleum source facies may be a part of a rock sequence containing, in addition, rocks which cannot be regarded as petroleum source rocks" and then continues: "The conditions for the development of the petroleum source mass containing a number of petroleum source facies can be identified as petroleum source formation." It is obvious from these words that facies are understood to be rock varieties and a petroleum source formation --

the depositional environment of a petroleum source mass.

However, I. O. Brod does not adhere to such an interpretation of the petroleum source formation and in other instances he defines it as a rock mass; this is similar to some instances where he interprets a facies as a depositional environment and then again as rocks [or] sediments. Thus, he writes: "Only those deposits in which the transformation of the organic matter leads to the bituminization can be called petroleum source facies." The petroleum source formation is only that rock mass or rock series in which, in the process of the molecular displacement, the stable products of the transformed organic matter originate and remain together with the mobile hydrocarbons" (p. 75). It is evident here that petroleum source facies and formations are understood to be rocks in relation to the processes which take place in the rock itself in the bituminization of organic matter and not in relation to the forming of the series in the process of sedimentation. Thus, the concept "facies" here loses to a considerable extent a paleogeographic significance and it reflects the processes related to the transformation of the accumulated organic matter into bitumen, and not the conditions of the accumulation of organic matter.

K. P. Kalitskii is quite right in believing that "in referring to a certain rock series as a petroliferous facies, we mean that the deposition of organic matter which participated in the petroleum generation in a suite proceeded simultaneously with the deposition of the suite itself" (1944, p. 209).

We think that the future use of the term facies ought to conform to the general concept of facies (geological facies). If our definition given above happens to be sufficient in providing the meaning and content which is proper for the concept, then it is advisable that individual characteristics of facies, which have been segregated or will be segregated according to some facies features or their complexes, be based on the same basic content and meaning which follows our definition of the concept "facies".

Inasmuch as certain special terms in designating various geological facies have received recognition in the literature, we suggest the following definitions of these terms.

"Geochemical facies" --geological facies segregated by geochemical facies features.

"Lithologic facies" --geological facies segregated by lithologic facies features.

"Carbonate facies" --geological facies of carbonate rocks which are segregated by geochemical as well as lithologic or paleontological facies features.

"Faunal facies" --geological facies segregated by faunal facies features.

"Floral facies" --geological facies segregated by floral facies features.

"Petroliferous facies"⁴ or better, "petroleum-source facies"--geological facies reflecting the conditions of the deposition of the source material in the generation of oil and gas.

"Coal-bearing facies"--geological facies reflecting the conditions of the accumulation and deposition of the source coal-forming material.

All facies of this type or, more accurately, the varieties of geological facies and their individual occurrences, reflect to a different degree the variety of individual specific depositional processes, and the conditions of their progress.

Thus, the facies volume is determined by those facies features, which are accepted as basic, guiding, or correlative for each specific case.

As methods of investigation and study of the sedimentary deposits become improved, and new facies features are determined, the variety of geological facies naturally will increase in number, and their volume will accordingly become more accurate.

We fully agree with L. V. Pustovalov that the main development of various rock-forming organisms "in one or another part of a rock unit enables us to segregate within it different facies, the names of which indicate the main development of one or another group of organic remains in the rock" (Pustovalov, 1940, pt. II, p. 279).

Facies can be segregated by different facies features, and in each given specific case, the facies must be segregated by features which are reliable and well expressed. To segregate facies means to recognize the features resulting from changes in time and space. In turn, it discloses broad possibilities for learning very important laws of these changes in solving theoretical and practical problems. A principal problem of the study of facies is the interpretation of facies features, that is, the understanding of those processes which brought about these features in a deposit or a rock. The more varied and detailed the studies of rocks and deposits and of their facies features, as well as of contemporary processes of sedimentation

⁴ We have quoted the term "petroliferous facies" because of its being widespread in the literature and it should be considered a synonym for the more exact term "petroleum-source facies".

the more accurate and broad will be knowledge of those processes which took place on the earth's surface during the past periods, under various conditions and environments.

This study, on the one hand, gives an opportunity for acquiring a better knowledge of the conditions and laws which lead to the formation of mineral deposits, of their paragenetic relationship to other sedimentary deposits, and methods for making the prospecting for mineral deposits more accurate, and, on the other hand, for a broader study of the history of the earth's development and the processes which took place in it.

Summarizing the above material, we can generally point out the three following tendencies in defining the facies concept:

I. The concept of "facies" is connected with sedimentary rocks (conditions of a totality of conditions of the formation of rocks or sediments; totality of characteristics of sedimentary rocks, of sediments; the types of rocks, sediments, etc., of a zone in the sedimentation area).

II. The concept of "facies" is not related to sedimentary rocks (totality of conditions on the surface of the earth's crust, surface sections, landscapes, facies of effusive rocks, etc.).

III. Non-uniform interpretation of facies (qual, triple, etc.).

Despite a great difference in interpretation and discussion of facies, types of rock or deposits are usually separated as facies in compiling maps of facies or facies sections. All other concepts, however widely they may be used in the literature, have not received a specific expression.

As for the rock types, in our opinion, they can hardly be interpreted as facies. Not every rock type is a facies. For instance, if limestone is changed to marble, both types of rock will constitute one facies. If a sand is changed to sandstone, they can not always be separated as facies, especially under similar conditions of sedimentation. Many such instances can be given here for conglomerate and gravel, clay and slate, etc.

And so, the idea of facies is conditioned by the removal from a rock of all changes not connected with the depositional process and, therefore, not reflecting the sedimentary environment and the character of the processes which took place during that period.

The segregation of facies is not the separation of rock types by any characteristic of a rock. Facies are segregated only by primary

facies features. The segregation of facies in deposits of modern basins is an entirely different matter. Here the sediments have still preserved all characteristic properties affected by the processes which accompany sedimentation and reflect the conditions of the sedimentary process.

CHAPTER V. SOME RESULTS OF THE ANALYSIS OF FACIES IN MIOCENE AND PLIOCENE DEPOSITS OF EASTERN GEORGIA

As an illustration of our concept of facies, let us take a brief look into the peculiarities in the distribution of geologic facies of Miocene and Pliocene deposits in the territory of eastern Georgia. The facies analysis of deposits of this territory has been considerably helpful in interpreting the trend of development of this region and in solution of various other complex geological questions.

The principle used in preparation of facies and thickness maps is not new. Maps depicting these features for the territory of Georgia have been made by earlier workers (Buleishvili, 1947; Varentsov, 1950; Khain and Shar-danov, 1952; Markevich, 1954; Varentsov and Mordovskii, 1954; and others). Our facies and thickness maps are based on published material as well as on a substantial amount of personal investigation. The facies analysis as carried out provided us with solution of many interesting and important problems of the geology of eastern Georgia (Markevich, 1954). However, in this case we are presenting for illustration only a small section of the history of geologic development in eastern Georgia.

The development and existence of the Kura depression between the large regions of the Great and Little Caucasus is naturally related to the same geotectonic processes which existed to some degree in the vast geosynclinal territory of the Caucasus (fig. 4).

Let us study some features of the history of the development of the Kura-Alazan depression during Upper Miocene and Pliocene times, as well as the distribution of facies in the basins of that period.

The configuration of the Lower Sarmatian basin and its sedimentary environment were in a general way inherited from older basins. (figs. 5, 6, 7),⁵ but the typical fauna which had

⁵The principle of drawing of facies section attached is based on the analysis of facies and deposit thicknesses, with the top of a given stratigraphic complex of deposit reduced to the horizontal plane.

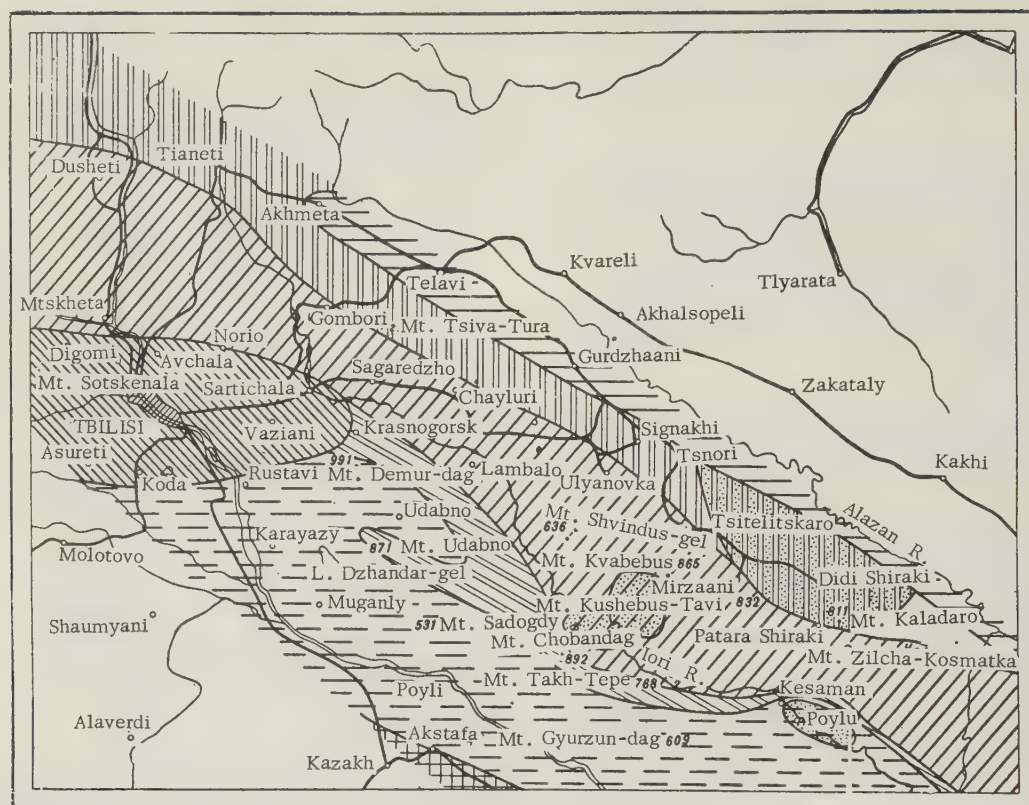


FIGURE 4. Map of tectonic zones of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

- | | | | |
|--|------------------------------------|--|--------------------------------------|
| | 1 - Alazan depression | | 5 - Kura-Kakhetian depression |
| | 2 - zone of the Kakhetian uplift | | 6 - Little Caucasus |
| | 3 - Iori depression | | 7 - Trialeti zone |
| | 4 - uplift zone of the Iori region | | 8 - areas of probable ancient uplift |

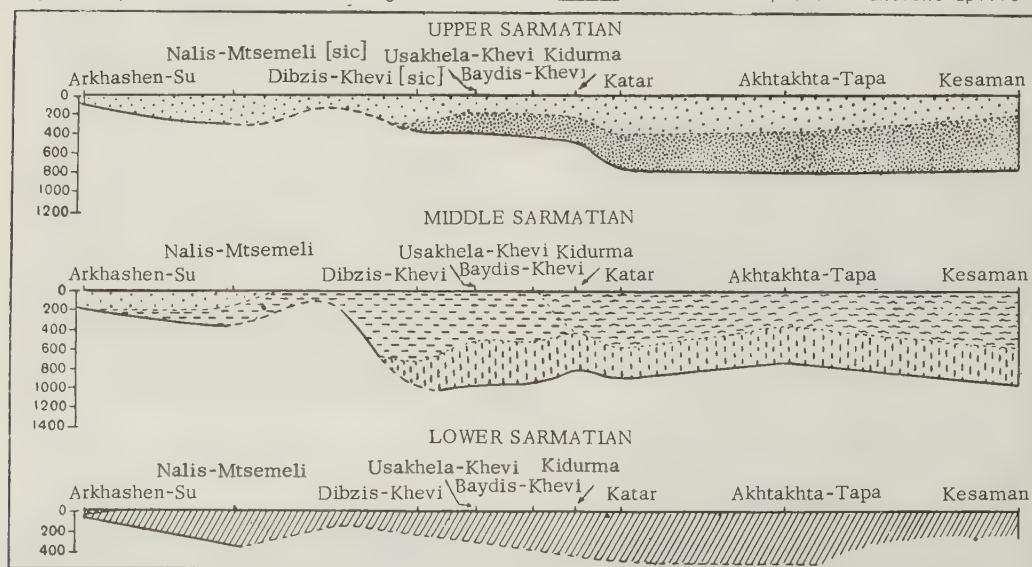


FIGURE 5. Facies section showing distribution of beds of the Sarmatian stage between Arkhashen-su and Kesaman (data from V. P. Markevich, 1954) (Legend continued on next page)

Legend for Figure 5.

- | | |
|--|---|
| 1 - variegated beds | 4 - beds of the Chobandag series |
| 2 - beds containing marine Upper Sarmatian fauna | 5 - beds containing <i>Cryptomactra pes anseris</i> |
| 3 - beds with typical Middle Sarmatian non- <i>Cryptomactra</i> fauna of southern Russian type | 6 - beds containing fauna of Lower Sarmatian age |
| | 7 - boundaries of facies |

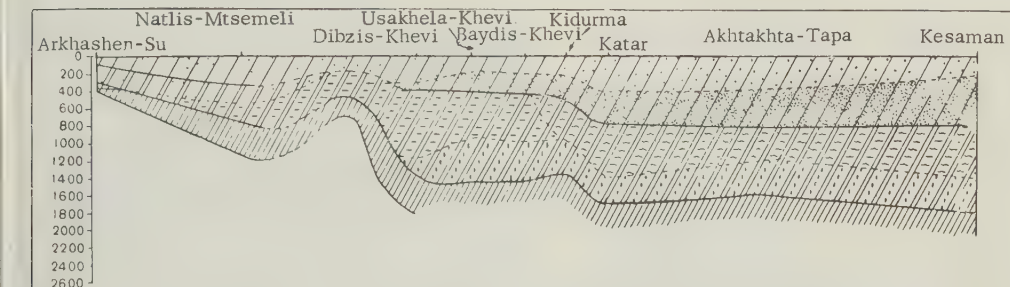


FIGURE 6. Facies section showing distribution of beds of the Sarmatian stage between Arkhashen-su and Kesaman (data from V. P. Markevich, 1954)

- | | |
|------------------------------|--|
| 1 - Upper Sarmatian | 7 - beds containing marine Upper Sarmatian fauna |
| 2 - Middle Sarmatian | 8 - beds containing typical Middle Sarmatian non- <i>Cryptomactra</i> fauna of the southern Russian type |
| 3 - Lower Sarmatian | 9 - beds of the Chobandag series |
| 4 - stratigraphic boundaries | 10 - beds containing <i>Cryptomactra pes anseris</i> |
| 5 - facies boundaries | |
| 6 - variegated beds | |

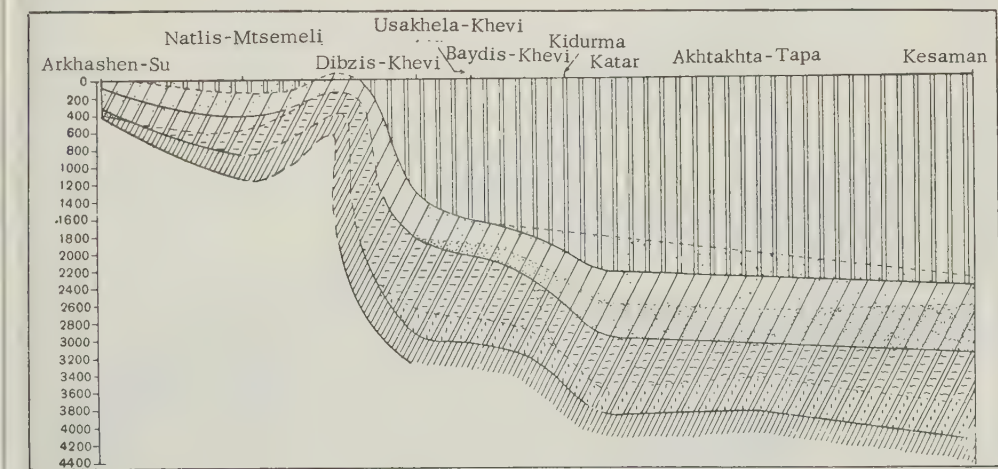


FIGURE 7. Facies section showing distribution of beds of the Sarmatian stage and the Shirakian beds at the end of the Shirakian period between Arkashen-su and Kesaman (data from V.P. Markevich, 1954)

- | | |
|--|--|
| 1 - Shirakian layer | 8 - beds containing typical Middle Sarmatian non- <i>Cryptomactra</i> fauna of the southern Russian type |
| 2 - Upper Sarmatian | 9 - beds of the Chobandag series |
| 3 - Middle Sarmatian | 10 - beds containing <i>Cryptomactra pes anseris</i> |
| 4 - Lower Sarmatian | 11 - boundaries of facies |
| 5 - stratigraphic boundaries | |
| 6 - variegated beds | |
| 7 - beds containing marine Upper Sarmatian fauna | |

occupied the basin up to that time disappeared, and new forms characteristic of the Lower Sarmatian basin made their appearance. The faunal change is evidently related to a change of the regimen of the basin: the depth during the deposition, water salinity, currents, climate, etc. (figs. 8 and 9).

Because of poor exposures of the Lower Sarmatian deposits, it is difficult to establish a detailed physico-geographic picture of the basin for that period. Exposures of Lower Sarmatian rocks and the underlying deposits are related to the belt of the Iori uplift (which at present is well expressed geomorphologically as a chain of ranges extended latitudinally along the right shore of the River Iori), or more accurately to the western part of this belt. To the east of the Chobandag range, within Georgia, no rocks more ancient than the Middle Sarmatian are exposed. The lower formation of the Sarmatian stage in the Iori belt extending from the Aram-Dar region eastward to the Akhtakhta-Tapa Mountain region is represented by a clay facies with very sparse

fauna. Still farther to the east, no exposed rocks of this age are present, and it seems that no particular changes in the belt can be observed still farther to the east. To the south and west of this belt the part played by sandy formations increases, with deposits showing increased fossil content.

Furthermore, to the south of the belt of the Iori uplift in the vicinity of the Sovkhoz Udabno (Monastyrskaya ridge) an increase in thickness of these deposits from 250-300 to 500 m and more is noted.

Complete replacement of the Lower Sarmatian deposits containing a marine fauna with variegated unfossiliferous deposits is noted south of the Sovkhoz Udabno and the Arkhashen-su ravine. The development of coarse-grained sandy deposits during the Lower Sarmatian age is also noted in western regions adjacent to the eastern subsidence of the Tbilisi-Malkhazovka tectonic shelf and in regions south of the Iori uplift. Thus, in the vicinity of the village Malkhazovka sandy rocks are predominant in

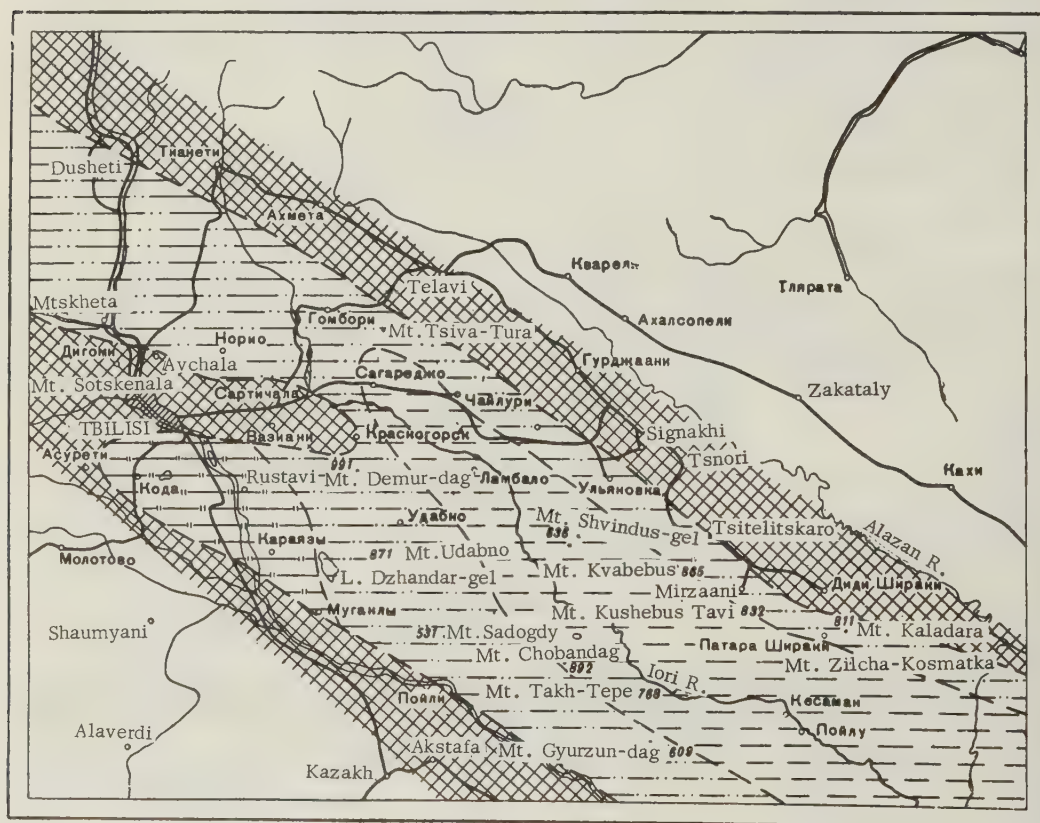







FIGURE 8. Map of facies of the Lower Sarmatian beds of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

- | | |
|---|--|
|  1 - land |  3 - clayey-marly beds |
|  2 - sandy-clayey beds with conglomerate interbedding |  4 - variegated beds |
| |  5 - boundaries of facies |

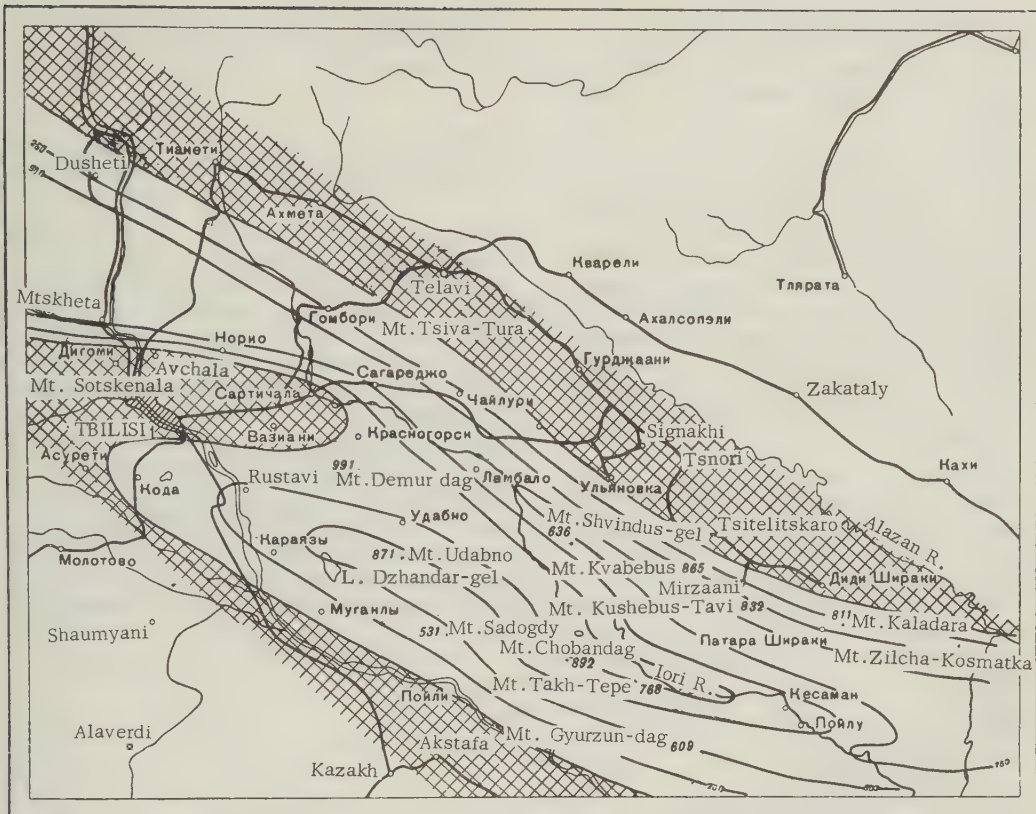


FIGURE 9. Map showing distribution of thicknesses of the Lower Sarmatian deposits in the Inter-mountain basin within eastern Georgia (data from V. P. Markevich, 1954)

1 - land

2 - isopachs at 250 m intervals

the section. Here, the Lower Sarmatian is represented by sandstone, sands, conglomerates, dolomitic limestones, and more rarely by clays. Here and there the section is very fossiliferous.

A similar picture is observed also in the vicinity of the Arkhashen-su ravine and in the western part of the Sovkhoz Udabno area where coarse-grained sand formations are dominant in the section. Thicknesses of these deposits in the area of the western part of the Udabno-Monastery ridge are greater than around the Arkhashen-su ravine. So, if the thickness of the Lower Sarmatian in the first region is about 300 m, then it is only 100-120 m in the second. Sandy formations play a significant part also in other regions south of the Iori uplift (along the small rivers Legvistskhali, Armutla, and others).

Thus, the influence of the Iori uplift on the distribution and thickness of facies in the basin are quite considerable. The northern boundary of the Lower Sarmatian basin or, more accurately, the northern boundary of the extension of Lower Sarmatian deposits is most probably

concealed under the so-called Orkhev overthrust, and the axis of the maximum depression of the Lower Sarmatian basin passes somewhere not far to the northeast of the villages of Gombori and Chailuri, since the increase of thickness and the clayish nature of the Lower Sarmatian is noted northeast from the village of Norio, and the thickness of these deposits in the vicinity of the village of Udzharmo approaches 500 m as compared with 80-100 m around Norio.

The Kura-Kakhetian zone (south of the Iori ridges) apparently is a depression of smaller depth, and the section indicates a considerable sandy deposit which increases in a southwesterly direction. Furthermore, in the western part of the Kura-Kakhetian depression (Karayazy Steppes) the deposition of variegated sediments devoid of marine fossils took place during the entire Lower Sarmatian period, with the influence of variegated facies also having been manifested, more or less, in more easterly regions within this zone.

Thus, in broad terms, the southern boundary of the Kura-Iori depression was inherited from

the preceding Konka-Karagan basin. The changes that took place then were related to replacement of principal elements of the fauna and, possibly, with rejuvenation of the sources of terrigenous material.

During the subsequent Middle Sarmatian period the basin underwent considerable facies changes. The bottom of the basin experienced a sizeable subsidence with its deepest part moving northward into the area of the southern slope of the present Kakhetian ridge, in which place the deposition of sediments was greatest, attaining, for instance, 2-3 km in the vicinity of the village of Gombori (data by I. E. Karstens), but to the south, in the area of the Chatma Valley, Valley, the thickness of these deposits drops to 1000 m (figs. 10-12).

It is very difficult to comment on the northern shoreline, but the large thickness of the Middle Sarmatian in the vicinity of the village Gombori leads us to conjecture that the north shore of the Middle Sarmatian basin was very steep and that at the present time some parts of the shore

deposits are hidden from view under older upturned and overthrust deposits associated with the Orkhev fault.

The first half of the Middle Sarmatian period was characterized by the development of various markedly different facies in the basin (fig. 10).

A deep sea part of the basin was surrounded by the *Cryptomactra* facies. In the western regions (Arkhashen-su, Udobno, and others) coarse-grained rocks, sandstones and even conglomerates, were developed, and farther west of these regions, within the Karayazy Steppes, variegated sediments were deposited. Remnants of normal marine faunas are absent in variegated deposits, and large accumulations of Middle Sarmatian shells of South Russian affinities are observed in sandy clay sediments that replace the variegated rocks toward the east. The change from variegated sediments to sediments containing marine faunas is complete within the Udobno region. Thus, whereas in the area of the Natlis Mtseveli monastery



FIGURE 10. Map of facies of the lower part of the Middle Sarmatian beds of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

1 - land

2 - beds containing typical Middle Sarmatian fauna of the southern Russian type





3 - beds containing *Cryptomactra pes anseris*

4 - variegated beds

5 - boundaries of facies



FIGURE 11. Map of facies of the upper part of the Middle Sarmatian beds of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

- | | |
|--|--|
|  1 - land |  4 - variegated beds |
|  2 - beds with typical Middle Sarmatian non- <i>Cryptomactra</i> fauna of southern Russian type | 5 - beds containing <i>Cryptomactra pes anseris</i> |
|  3 - beds of the Chobandag series | 6 - boundaries of facies |

The upper part of the Middle Sarmatian is not represented by a variegated facies, near the Mountain Udobno (Garedzhinskii monastery) the whole section of the Middle Sarmatian is represented by sandstone containing rich Middle Sarmatian faunas; here variegated sediments are completely absent. Truly, such an assertion is not entirely accurate when considering that the lower part of the conformable variegated bed covering the Middle Sarmatian could have formed at the end of the Middle Sarmatian period as well. But this point is impossible to establish, and actually, it is not vitally important, because the replacement in an eastern direction of variegated deposits by sediments containing a marine fauna is clearly exposed and cannot give rise to doubt.

Thus, during the first half of the Sarmatian period, deposition of variegated sediments took place at the extreme southwest of the Kakhketian plain in the vicinity of the Karayazy Steppes; these sediments are replaced by sediments

containing a normal marine fauna in a westward direction in the vicinity of the Natlis-Mtsemeli monastery. However, *Cryptomactra pes anseris*, which in most places is characteristic of the lower part of the Middle Sarmatian, is absent in the fauna. Shells of *Cryptomactra pes anseris* occur in small numbers in the section of the Middle Sarmatian in the vicinity of Dibzis-Khevi and east of it. In addition to *Cryptomactra pes anseris*, in the midst of a varied Middle Sarmatian fauna, in the vicinity of Dibzis-Khevi, we have detected huge shells of *Mactra fabreana* which in most places are characteristic of the upper part of the Middle Sarmatian. All this points to the replacement of a shallow-water by a deep-sea environment in the vicinity of Dibzis-Khevi, and, therefore, explains the joint presence here of *Cryptomactra pes anseris* and *Mactra fabreana*.

Farther to the east the *Cryptomactra* deposits occur everywhere in the lower part of the Middle Sarmatian section. In the northern

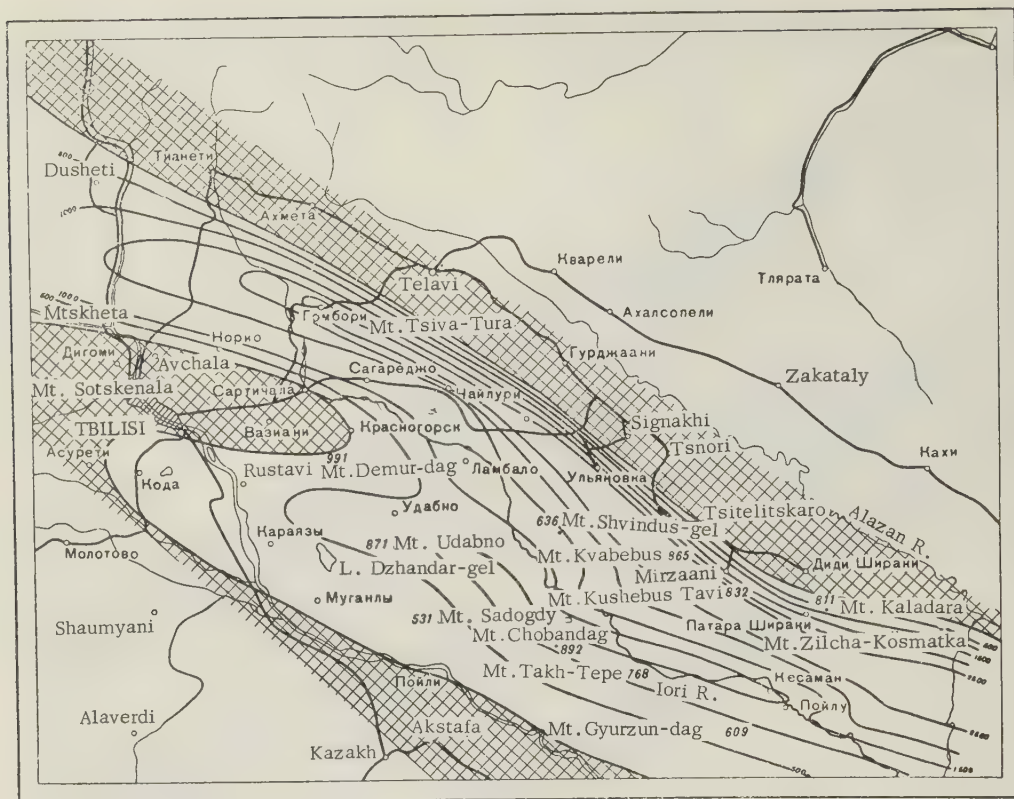


FIGURE 12. Map showing distribution of thicknesses of the Middle Sarmatian beds in the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954).

1 - land

2 - isopachs at 500 m intervals.

direction faunas within *Cryptomactra strata* are present in the area of Chailuris-Khevi and other areas.

Thus, the first half of the Middle Sarmatian period is characterized by the development in the Kura-Iori basin of three large facies complexes of deposits, with the shore and shallow-water deposits being expressed by two definite complexes: variegated sediments which are developed only in areas of the southwestern and of the Kakhetian plain, and sandy-clay deposits widespread in the shore and shallow-water parts of the basin.

During the second half of the Middle Sarmatian period, facies differences become even more accentuated. In addition to the facies complexes described above, one more complex of sediments clearly different from the others makes its appearance under the name of Chobandag suite (fig. 11).

The influence of the variegated facies increased and extended somewhat farther to the east, replacing progressively more recent deposits of the Middle Sarmatian. Deposits containing typical Middle Sarmatian fauna replace deposits of the *Cryptomactra* facies over

a sizeable part of the territory of South Kakheti, spreading nearly across the entire basin of the Middle Sarmatian Sea, in some southeastern areas of South Kakheti. From the vicinity of the Baidis-Khevi ravine and the middle part of the Chobandag ridge toward the east, distinctive, originally basically clay deposits of the Chobandag suite were developed. Sparse faunas found in the beds indicate their Middle Sarmatian age. The development of the Chobandag beds in all probability did not end in the Middle Sarmatian and extended into the beginning of the Upper Sarmatian period.

In the southeast (Palan-Tiukian) and possibly to the east, deposition of the *Cryptomactra* facies, which started at the beginning of the Middle Sarmatian period, continued during the period of the Chobandag beds.

In northern areas adjacent to the foot of the Kakhetian ridge, no trace of the Chobandag facies has been observed. Thus, the facies of the Chobandag suite is characteristic only at the southern part of the basin, and deposition of this facies was related to the southern sources of the sediments.

The greatest thicknesses of the Middle Sarmatian

an deposits (fig. 12), up to 2-3 km are observed in northern areas. This indicates that the greatest sag of the Middle Sarmatian basin is located in the northern part of South Kakheti in the vicinity of the villages of Gombori and Nailuri. The effect of the Iori uplift during the Middle Sarmatian period on the distribution of facies in the basin was considerable; particularly, this effect is evident in the nature of the hobandag beds and in the content and distribution of the variegated deposits. The development of deposits of the southern part of the basin, on the whole, depended on sources of terrigenous material derived from the Little Caucasus, while the northern more depressed and vast territory of the Kura-Iori depression was fed with detrital material from sources located in the region of the Great Caucasus. However, it should be noted that in the Middle Sarmatian period, the basin experienced a comparatively great subsidence, and it is possible that its boundaries were somewhat extended both north and south.

Such a hypothesis is based on the more





widespread areal occurrence of Middle Sarmatian deposits and their considerable thickness. Moreover, considering the *Cryptomactra* facies of the Middle Sarmatian to be of relatively deep-water type, the occurrence of *Cryptomactra* strata in the vicinity of Chailuris-Khevi indicates the location of the shore of the Middle Sarmatian basin to be much farther to the north of this area. In this case, also, the theory expressed earlier (V. P. Markevich, 1954) substantiates the tectonic overthrusting of more recent strata by older deposits overturned and thrust from the north. Therefore, north of the Orkhev thrust line the rocks of this age do not occur on the surface.

The Upper Sarmatian period is characterized by some decrease in basin size and by a considerable freshening of its water (figs. 13-15).

If during the first half of the Upper Sarmatian the variegated facies were concentrated basically in the southwestern part of the plains of Kakheti, and did not extend beyond the Iori



FIGURE 13. Map of facies of the lower part of the Upper Sarmatian beds of the Intermountain basin within eastern Georgia (data from V.P. Markevich, 1954).

- | | |
|--|--|
|  1 - land |  3 - variegated beds |
|  2 - sandy-clayey beds containing marine mactra |  4 - boundaries of facies |

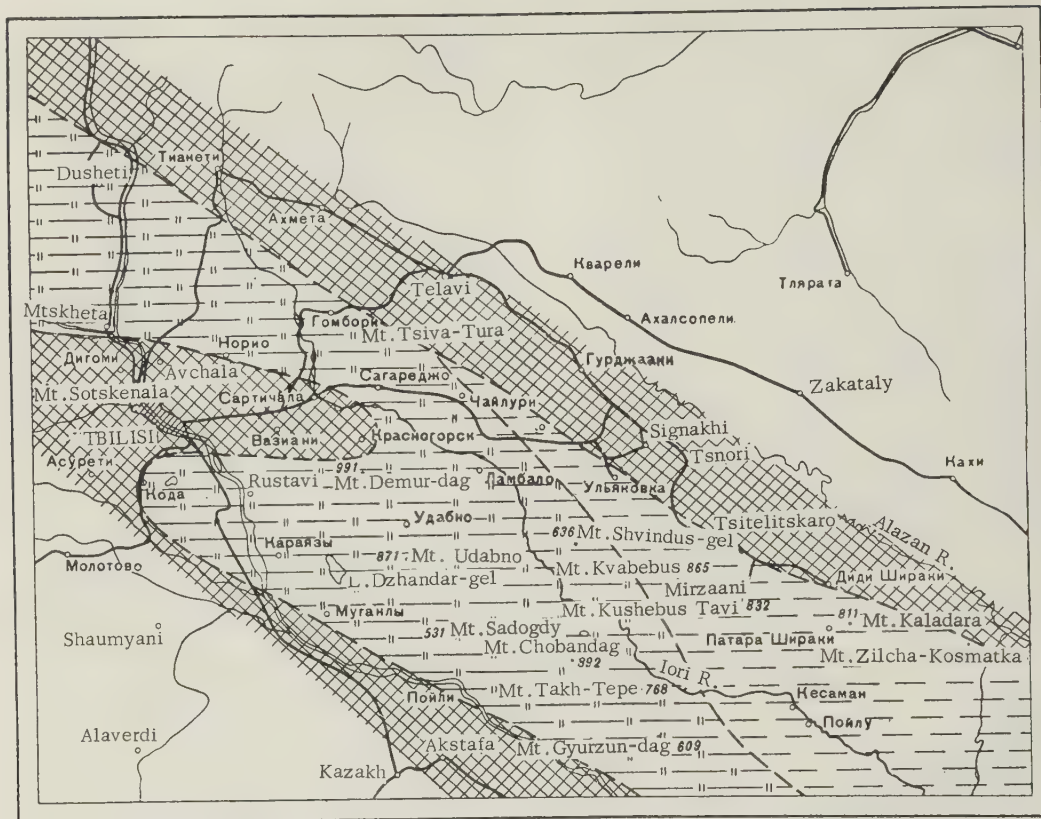






FIGURE 14. Map of facies of the upper part of the Upper Sarmatian beds of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954).

-  1- land
 2- variegated beds lacking marine fauna
 3 - variegated beds containing Upper Sarmatian macrofauna
 4 - boundaries of facies

uplift, then in the second half of the period, the effect of these facies was extended almost throughout the depression (figs. 5-7).

In the adjacent area west of the Kartalin depression, judging from works by M. I. Varentsov (1950) and others, considerable freshening took place in the basin at the very beginning of the Upper Sarmatian, and, possibly, also at the end of the Middle Sarmatian.

Relatively normal marine conditions during the first half of the Upper Sarmatian period existed only in South Kakhetiya except in western areas. Thus, in comparison with deposits of the Middle Sarmatian Sea, the Kura-Iori basin of the Upper Sarmatian differed by a lesser variety of facies complexes. Essentially, two different categories of deposits were developing during the Upper Sarmatian period: variegated sediments and normal marine types containing marine faunas.

The development of variegated deposits (fig. 7) took a long time in the southeastern part of the Kakhetiya Steppe, but farther east-

ward conditions necessary for the development of variegated rocks were less effective, ultimately resulting in a gradation of these sediments into normal marine deposits containing marine faunas. The absence of variegated deposits in sections of Neogene sediments in the Kakheti region farther to the north, in the area of the southern slopes of the Kakhetian ridge, furnishes evidence that the variegated facies was developed in the southern part of the area investigated, during that period exclusively, and therefore, these deposits are characteristic only of the southern part of Neogene basins.

One may also note that the extent of the variegated facies increased with the passing of time to more recent deposits of the Neogene. At the end of the Upper Sarmatian period large movements apparently took place in the area of the Great Caucasus, and those were related to the appearance of new, thick sources of terrigenous deposits and to a considerable sagging of the Iori depression in the area between the Iori uplift and the Gorno-Kakhetian Ridge. But the south Kura-Kakhetian part of

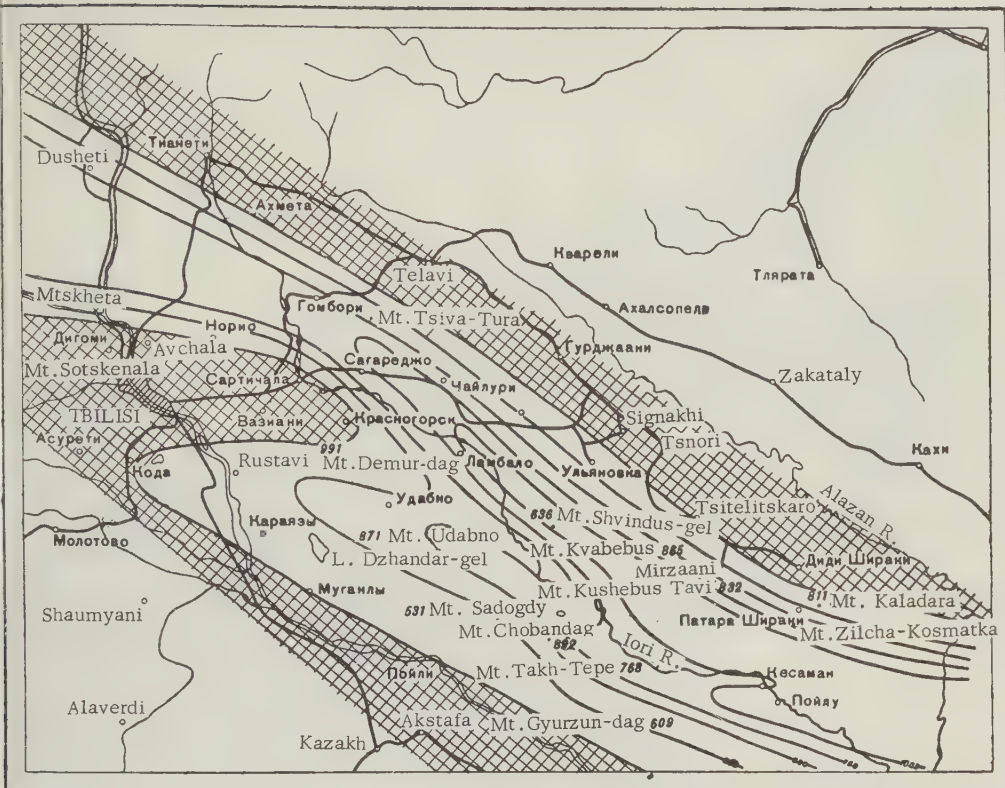


FIGURE 15. Map showing distribution of thicknesses of the Upper Sarmatian bed in the Inter-mountain basin within eastern Georgia (data from V. P. Markevich, 1954).

the depression, farther to the south of the uplift, did not undergo any serious subsidence. The process of deposition of variegated sediments of small thickness continued in this part of the basin (figs. 7, 16-18).

As for the Iori sag, during the entire period of deposition of the Shiraki beds a rapid sagging of the basin bottom took place, and the amount of subsidence was very large; this may be estimated simply on the basis of coarse-grained deposits having accumulated to over 1000 m during this period. If during this time of deposition of the Shiraki beds, conglomerate deposits were accumulating along the main boundary of the northern basin, then with increasing distance from the shore southward, the conglomerate deposits were replaced with sandy and, in the southernmost Iori areas, with argillaceous sediments.

In the Udabno region the thick (about 500 m) sandy deposits are usually referred to the Shiraki beds; but this assumption is hardly admissible. Firstly, nowhere in any area south of the Iori uplift are any deposits similar to facies of the Shiraki beds observed. Everywhere here variegated sediments overlain by the Akchagyl beds. Secondly, sediments

in the Udabno area usually referred to the Shiraki beds are characterized by prevalence of coarse-grained sandy deposits, a fact which distinguishes these sediments from the sandy-clayey section of the Shiraki beds in adjacent northern areas. Finally, the mineralogical composition of these sediments is quite different from that of the sediments of the Shiraki beds of adjacent areas (K. G. Chubinshvili, 1948). During the development of the Shiraki beds the part played by the zone of the Iori uplift in the facies distribution in the basin was significant, and the zone prevented the merging of facies of the northern and southern areas of the basin, though at times variegated deposits of the southern zone penetrated this barrier in the northern zone, i. e., into the Iori sag; this, however, indicates the presence of variegated sediments in the upper part of the Shiraki beds in the vicinity of the natural boundary of Kila-Kupra, and it gave the sediments a blotchy appearance. Thus, the above-mentioned sandy layer in the vicinity of Udabno should be referred to deposits of the Akchagyl period and not to the Shiraki beds, and the analogues of the latter for the entire territory south of the Iori uplift should be looked for in the upper layers of variegated deposits which usually belong to the Upper Sarmatian (Eldar suite).



FIGURE 16

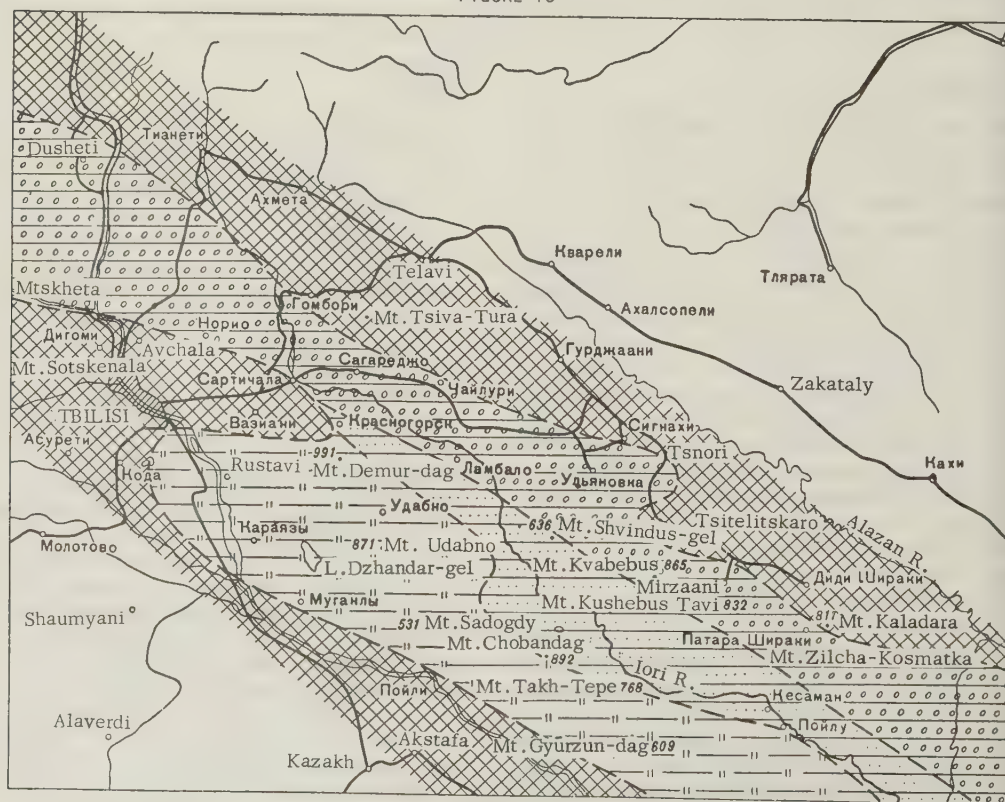


FIGURE 17

FIGURE 16. Map of facies of the lower part of the Shirakian beds of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)



FIGURE 17. Map of facies of the upper part of the Shirakian beds of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

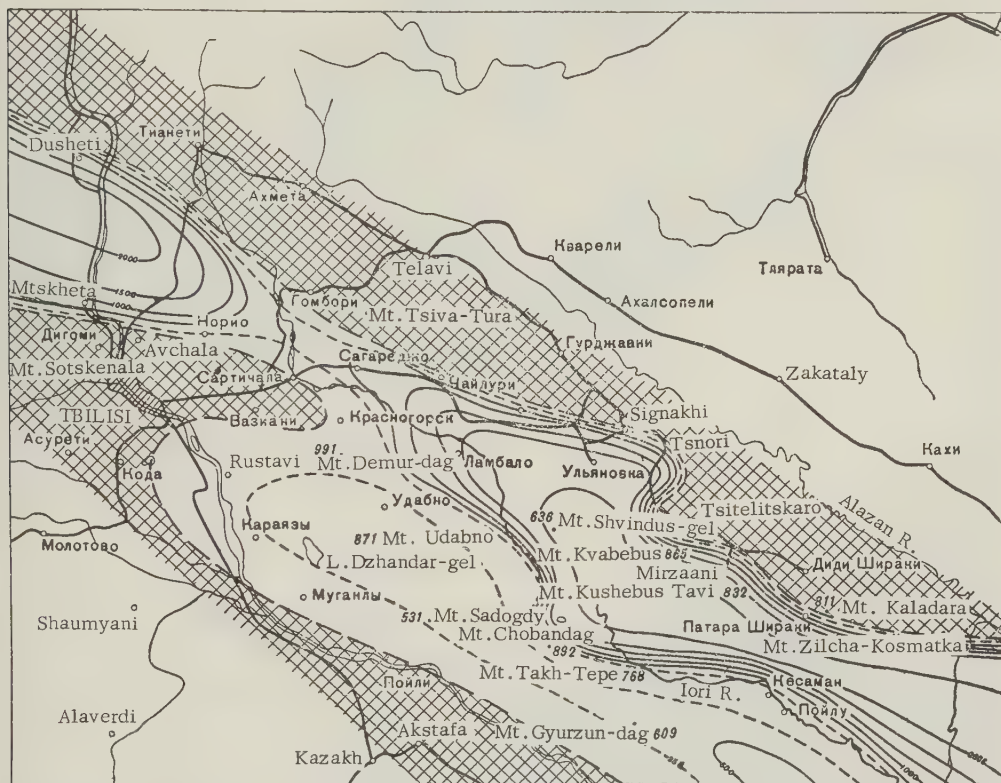
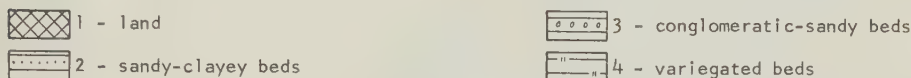
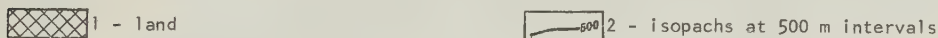


FIGURE 18. Map showing distribution of thicknesses of the Shirakian beds in the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)



At the end of the period of the deposition of the Shiraki beds, the entire Kakhetiya area experienced considerable tectonic activity followed by a partial disruption of former structures and finally by a general subsidence. Even if the basins after the Middle Sarmatian period decreased in size, then during the pre-Akchagyl period, following interruption in depositional and erosional processes which led to partial destruction of the existing structures, the Akchagyl transgression covered a substantial part of Kakhetiya, extending to some areas which earlier had been the sources of terrigenous material. Particu-

larly, the Akchagyl Sea advanced far to the north, its water covering the eastern end of the Gorno-Kakhetian ridge and the Alazan Valley region. Limestones in the vicinity of the Krasnye Kolodtzy (Red Wells) wedged out from under the level of the Akchagyl Sea as islands, and provided clastic material to the adjoining marine section (figs. 19-22).

In the southern area a sizeable portion of the Iori uplift was covered by the Akchagyl Sea. This portion probably has already been partially exposed during the Shiraki period and consider-



FIGURE 19

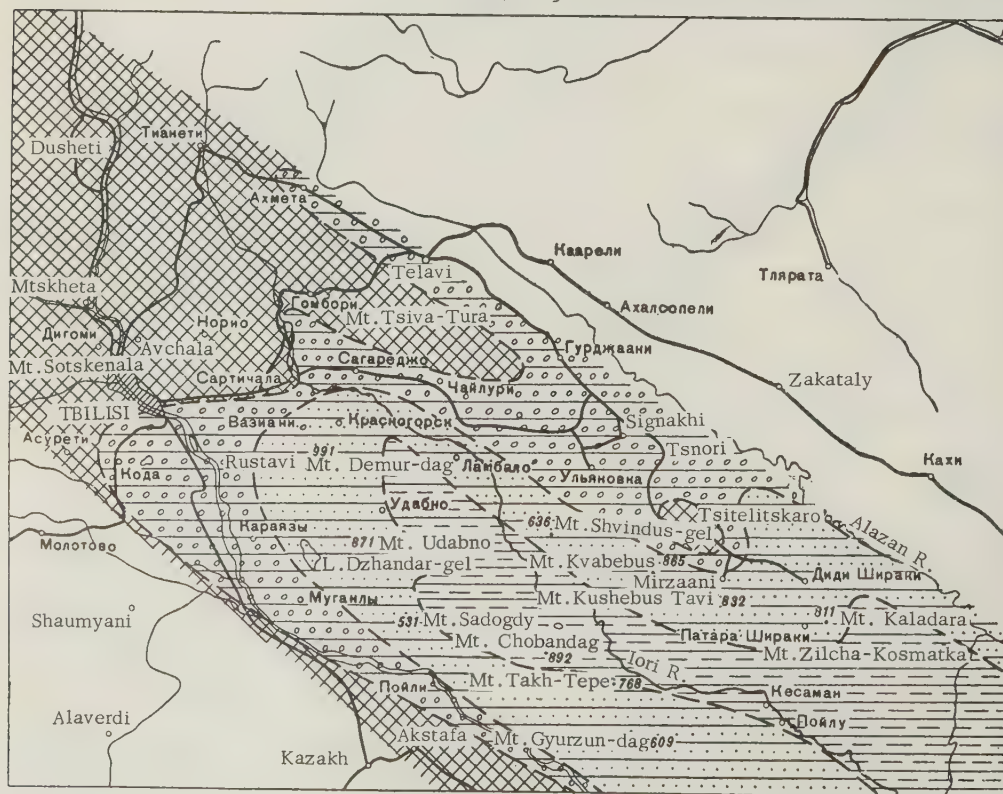


FIGURE 20

FIGURE 19. Map showing distribution of facies in the Akchagyl formation of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

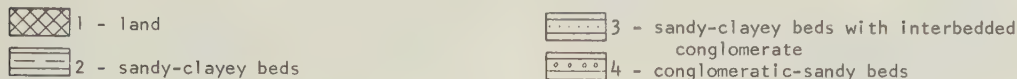


FIGURE 20. Map showing distribution of facies in the Apsheon formation of the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)

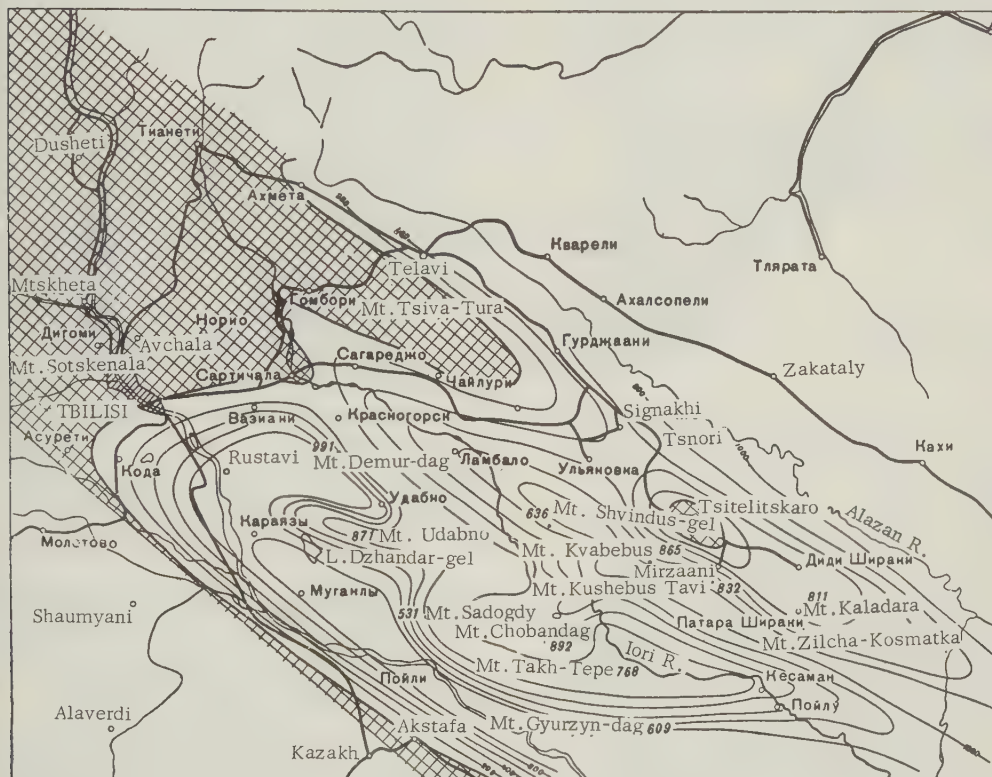
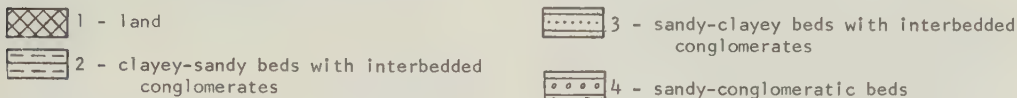


FIGURE 21. Map showing distribution of thicknesses of the Upper Pliocene beds (Akchagyl + Apsheon) in the Intermountain basin within eastern Georgia (data from V. P. Markevich, 1954)



ably disturbed during the pre-Akchagyl period. Apparently, at times a narrow belt of the uplift cropped out from under the Akchagyl Sea in the form of a bar.

Westward, the sea advanced and inundated earlier sedimentary deposits of the Tbilisi region. In a southern direction, the boundary line of the Akchagyl Sea also extended considerably to the south of the valley of the River Kura.

The increase in thickness of the Akchagyl deposits from west to east to 1, 000 m and more

indicates that the sagging of the depression progressed westward. Despite the wide extent of the Akchagyl Sea within Kakhetiya, its waters, most probably, did not advance into the adjoining Kartalin region, and the seashore line passed somewhere else in the Tbilisi area.

Subsequently, the vast Akchagyl Sea within Kakhetiya began to shrink from the periphery toward the center and to retreat toward the east. This eastward retreat of the basin continued in the Apsheon period when the normal marine environments existed only in the easternmost



FIGURE 22. Map showing distribution of marine fauna in the Upper Pliocene basin within eastern Georgia (data from V. P. Markevich, 1954)

- | | |
|---|--|
| <p>1 - land</p> <p>2 - zone of marine fauna penetration in the Akchagyl basin</p> | <p>3 - zone of marine fauna penetration in the Apsheeron basin</p> <p>4 - zone of maximum extent of the Akchagyl transgression</p> |
|---|--|

part of South Kakhetiya. Toward the end of the Apsheeron period the basin retreated still farther to the east beyond [the boundaries of] Georgia, and the territory of Kakhetiya was not subjected to transgressions any more.

The above factual material on the history and development of the Kura-Alazan depression and on the facies distribution in the basins shows abrupt changes of facies and of basin environments during the Upper Miocene and Pliocene periods.

If during the Lower Sarmatian period the basin was not characterized by facies varieties, then at the subsequent Middle Sarmatian period as a result of tectonic movements, changes of depths, and of basin outline, we see quite a complicated picture of the development of very distinct facies complexes: sediments characterized by the fauna *Cryptomacra pes anseris*, sediments of the Chobandag suite, sediments containing southern Russian-type faunas, variegated sediments, and others. The first three facies complexes make their appearance in the Middle Sarmatian, while the facies of variegated

sediments originated much earlier in the southwestern part of the basin.

In the subsequent Upper Sarmatian period, the basin becomes somewhat simplified and reduced in size, the complexes mentioned above disappear, and the variegated sediments develop on a large scale. During the Pliocene period, in connection with renewed tectonic movement in the north, the basin again becomes complex, new integrated deposits in the Shiraki beds characterized by sharp facies varieties make their appearance, etc. Thus, we see that during a short time the deposits of the basins underwent sharp facies changes caused by abrupt changes in sedimentary environment.

This question has been analyzed at greater length in an earlier paper (Markevich, 1954). Now, however, an attempt has been made to acquaint the reader with some principles of developing facies maps and facies sections from the material obtained by the author as a result of a study of this geologically complex area, and also the material of other investigators (D. A. Buleishvili, M. I. Varentsov,

7. E. Khain, and others).

Naturally, the development of maps is not limited to the existing facies maps and facies sections. If the maps of facies and facies sections show large facies units, then subsequent investigations can contribute to compilation of more detailed facies maps on which facies are separated by their distinct facies features; this naturally will enrich considerably our knowledge of the geologic development of the region and its oil and gas potential.

The basic pattern in the geological features of the area discussed above were determined precisely from the facies analysis of sediments forming the Kura-Alazan depression. Only by analyzing the distribution of various facies in time and space, by studying the facies features of stratigraphic sections of single areas, by compiling maps of facies and strata, as well as by constructing facies sections, was it possible to a considerable degree to interpret the geologic history of the regional development and to explain the highly intricate distribution of tectonic elements within the area. Thanks entirely to facies analysis the existence of an ancient uplift at the time of the development of Neogene sedimentary deposits and the existence of specific structures called "dynamic structures" were successfully established and ascertained (Markevich, 1954). Facies analyses also helped to understand problems of the stratigraphy of Miocene sediments, the distribution of various sediments in time and space, and particularly, the question of development of deposits of variegated strata and the Shiraki suite, as well as many other questions.

It seems to us that facies analysis will be widely used in the near future in geological investigations and will furnish the solution of many geological questions which have appeared puzzling for a long time.

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SOLUBILITY OF GYPSUM IN AQUEOUS SOLUTIONS OF SALTS¹

by E. B. Shternina

• translated by A. J. Shneiderov •

ABSTRACT

The system $\text{CaSO}_4\text{-NaCl-H}_2\text{O-CO}_2$, part of the system $\text{CaCO}_3\text{-CaSO}_4\text{-NaCl-H}_2\text{O-CO}_2$, was investigated through a range of NaCl concentrations. Calcite and gypsum solubilities were studied under conditions approximating those of the natural environment as closely as possible, for the particular purpose of determining the behavior of slightly soluble salts in the presence of readily soluble salts. In that even their saturated solutions remain dilute, concentration of calcite and gypsum is insignificant compared to that of NaCl. From a moderate amount of experimental data, then, it was possible to compute the solubility diagram for a complex system through application of the equations of Debye and Hückel. The Debye-Hückel approximate equation was found to be best suited to the purposes of this experiment. Solubility of gypsum in aqueous NaCl solutions (up to saturation) was calculated according to these equations. If a system goes to equilibrium very slowly, which is the case here, investigation is much simplified when it is possible to find the solubility by calculation. Mean effective ion diameters were assumed to be a function of concentration. A linear equation was obtained for the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$, to describe the relation between effective ion diameters and NaCl content. A method of calculation to determine gypsum solubility was verified for several similar systems according to experimental data from various authors. An approximate form is given for experimental linear equations for correlation of effective ion diameters with concentrations. It is possible that crystallographic dimensions of ions as well as solubility of added salts, constitute bases for an attempt to explain the complexity of dependence of the mean effective ion diameters upon concentration of systems that are composed of gypsum and aqueous solutions of readily soluble salts with no ions common to gypsum or ability to react chemically with it. -- D. D. Fisher.

We studied the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O-CO}_2$, a portion of the system $\text{CaCO}_3\text{-CaSO}_4\text{-NaCl-H}_2\text{O-CO}_2$, through the range of moderate concentrations of table salt [1]. Studies on solubilities of calcite and gypsum under conditions approximating those found in nature, are of interest not only from the practical point of view, but theoretically as well, in relation to the study on behavior of sparingly soluble salts in the presence of readily soluble salts.

A colorless and transparent crystal of natural gypsum from Karaul-Khang, Uzbek SSR, was used. An analysis showed the following calcium, sulfate, and water content to approach theoretical values; and, magnesium and chlorine to be absent:

	Ca	SO ₄	H ₂ O	Total
Theoretical content in gypsum (in percent)	23.28	55.79	20.93	100.00
Analysis results for gypsum from Karaul-Khang (in percent)	23.29	55.85	20.78	99.92

Chemically pure sodium chloride was used. It was recrystallized from a solution through which hydrogen chloride had been bubbled. So-called electrolytic water, doubly distilled through a quartz cooler, was used.

Solubility experiments were made in pyrex containers of 250 ml capacity; these differ from the usual containers only by having an outlet tube fused on to the bottom, making it possible to bubble carbon dioxide through the container. Partial pressure of carbon dioxide did not exceed one atmosphere. Gypsum was cut by a knife into small flakes along the cleavage plane because it loses water if crushed. About 10 gm of gypsum was placed in each container, then water or table salt solution was added. The loaded containers were placed in thermostated water bath that maintained a constant temperature of $25 \pm 0.05^\circ\text{C}$. The point of equilibrium in the system was determined by liquid-phase analyses, samples were taken regularly every 2-3 days. Mixing was discontinued 12 hours before the sample was taken; the solid phase settled, and the sample of completely transparent liquid was taken by pipette through a cotton filter.

The samples taken were analyzed for calcium, sulfate, and chloride. The latter was determined gravimetrically by precipitation with silver nitrate; the result was calculated as sodium chloride. Calcium was precipitated by ammonium oxalate, and weighed as the monohydrate of calcium oxalate.

¹ Translated from *Rastvorimost gipsa v vodnykh rastvorakh* sole: Akademiya Nauk SSSR, *Izvestiya Sektora Fiziko-khimicheskogo Analiza*, v. 17, 1947, p. 351-369.

The analysis results for Ca and SO₄ usually corroborated each other and served as a double check.

Equilibrium was considered to be established when analyses results became uniform; that is, when a difference in the CaSO₄ content per liter of solution in two consecutive samples was less than 0.01 gm. This required 7 to 10 days.

The solid phase present was studied at the crystallochemical laboratory of IONKh. Invariably it was gypsum, in agreement with Foret's [3] data.

Carbon dioxide has almost no effect on the solubility of gypsum in water and in table-salt solutions. Therefore, no consideration will be given in this paper to the presence of CO₂.

The systems studied, calcite and gypsum, are peculiar in the fact that even their saturated solutions remain dilute; in saturated solutions, concentration of these sparingly soluble salts are insignificant compared to the solubility of sodium chloride. This peculiarity provides a possibility for applying certain quantitative laws of the modern theory of solutions.

Application of the activity theory and equations of Debye and Hückel allows one, in some cases, to compute the solubility diagram of a complex system from a moderate amount of experimental data. This circumstance is important when a long period is required for an equilibrium to be established; for instance, dissolution of calcium carbonate. Following is the description of an attempt, using the above mentioned laws, in the studies of gypsum solubility in the presence of other readily soluble salts.

It is known [4] that the following simple relationship exists between activity and concentration

$$a = fx, \quad (1)$$

where a is activity, f is the activity coefficient, and x is the concentration.

Values for the activity can be obtained from experimental data on the solubility of gypsum in pure water and in dilute salt solutions by extrapolation of the solution concentration to zero.

The activity curve becomes linear and can be extrapolated easily to the intersection with zero ordinate, if the content of gypsum in the solution were plotted as square root of its ionic strength and not as the ordinary concentration of the solution. Thus obtained, the activity of gypsum in saturated solutions when table salt is present, is 11.0×10^{-5} molal; this corresponds to 0.830 gm CaSO₄ per liter

of solution (figure 1, table 1). The activity

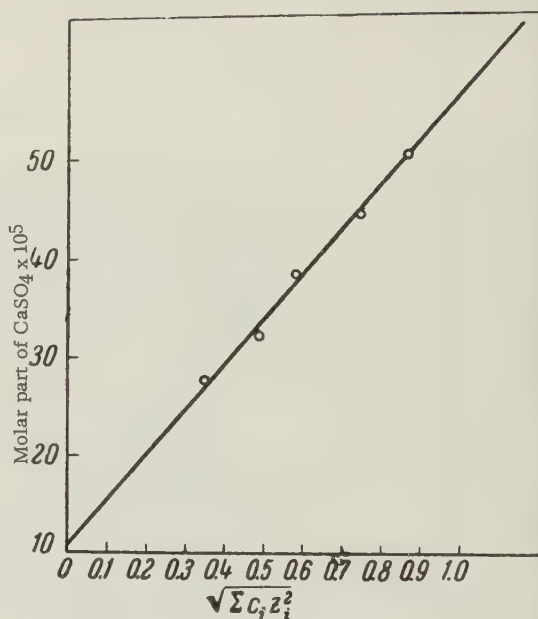


FIGURE 1. Determination of the activity of CaSO₄ in NaCl solution.

coefficient can be calculated from the Debye-Hückel equation (5):

$$-\ln f_{\pm} = \frac{\epsilon^2 z_1 z_2}{2DkT} \frac{x}{1 + \kappa b}, \quad (2)$$

where f_{\pm} is the mean activity coefficient of

gypsum and equal to $\sqrt{f_{Ca} f_{SO_4}}$ (the sign \pm will hereafter be omitted, although the mean ionic-activity coefficient of calcium and sulfate will be meant in all cases); ϵ is the electronic charge; z_1 and z_2 are the valences of calcium and sulphate; D is the dielectric constant of water; k is Boltzmann's constant; T is the absolute temperature;

$$\kappa = \sqrt{\frac{4\pi\epsilon}{DkT} \frac{N}{1000} \sum c_i z_i^2}; \quad N \text{ is}$$

Avogadro's number; $\sum c_i z_i^2$ is the ionic strength of solution; c_i is the number of moles of the i -th type of ion per liter of the solution; and b is the mean effective diameter of ions in solution.

By substituting numerical values for the constants [6] and changing from natural to common logarithms, the following expression is obtained for the 2-2-valent electrolyte (such as gypsum) at 25°C:

$$-\lg f = \frac{0.3601 \cdot 4 \sqrt{\sum c_i z_i^2}}{1 + 0.2325 b \sqrt{\sum c_i z_i^2}} \quad (3)$$

TABLE 1. Experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-NaCl-CO}_2\text{-H}_2\text{O}$ at 25° C. (E.B. Shternina and E.V. Frolova).

Point number	NaCl		CaSO ₄			$d_{\text{eff}}^{\text{NaCl}}$	$\sum c_i z_i^2$	$\sqrt{\sum c_i z_i^2}$	$\sum c_i z_i^2$	$\sqrt{\sum c_i z_i^2}$	$f = \frac{a}{x}$
	g/L	mol/L	g/L	mol/L	Molal part 10 ⁵						
1	0	0	2.083	0.01530	27.63	0.9993	0		0.1224	0.3499	0.398
2	2.488	0.043	2.438	0.01791	32.35	1.0013	0.086		0.2293	0.4790	0.340
3	4.972	0.085	2.899	0.02129	38.45	1.0034	0.170		0.3402	0.5834	0.286
4	10.072	0.172	3.405	0.02501	45.16	1.0077	0.344		0.5448	0.7380	0.244
5	14.870	0.254	3.850	0.02828	51.07	1.0114	0.508		0.7342	0.8570	0.215
6	19.807	0.339	4.312	0.03167	57.18	1.0153	0.678	0.823	0.9314	0.9651	0.192
7	50.035	0.856	6.003	0.04409	79.58	1.0379	1.712	1.308			0.138
8	86.103	1.473	6.865	0.05043	91.26	1.0612	2.946	1.716			0.121
9	119.867	2.051	7.216	0.05300	96.01	1.0837	4.102	2.025			0.115
10	138.766	2.374	7.236	0.05315	96.39	1.0958	4.748	2.179			0.114
11	152.134	2.603	7.230	0.05311	96.34	1.1049	5.206	2.282			0.114
12	168.204	2.877	7.161	0.05260	95.58	1.1140	5.754	2.399			0.115
13	180.746	3.092	7.056	0.05183	94.30	1.1214	6.184	2.487			0.117
14	231.513	3.961	6.654	0.04888	89.47	1.1501	7.922	2.815			0.123
15	283.495	4.850	6.102	0.04482	82.42	1.1812	9.700	3.114			0.133
16	317.228	5.427	5.816	0.04272	78.78	1.2014	10.854	3.295			0.140

and the so-called Debye-Hückel limiting law

$$-\lg f = 0.3601 \cdot 4 \sqrt{\sum c_i z_i^2} \quad (4)$$

for the most dilute solutions.

In equation (3) the mean effective diameter of ions in solution is expressed in Angstroms.

Our experimental data on gypsum solubility in aqueous solutions of table salt and carbon dioxide at 25° C are shown in Table 1 and Figure 2. In general, the data are similar to those of other authors [7], obtained at 20° C in the absence of carbon dioxide. This results from the fact that the temperature coefficients of solubility for gypsum and table salt are insignificant, and carbon dioxide has almost no effect on gypsum solubility. Activity coefficient of gypsum in saturated solutions can easily be obtained from these data using equation (1); it is plotted as a function of the square root of ionic strength and shown in Figure 2.

For dilute solutions, ionic strength was calculated according to the total concentration of sodium chloride and calcium sulfate. Beginning with approximately 5 percent sodium chloride solution (point 7 in table 1), however, ionic strength was calculated from the sodium chloride content alone; dissolved calcium sulfate was ignored because its concentration is insignificant compared with that of table salt.

To test the applicability of Debye-Hückel equations to our system, the activity coefficient for gypsum in solution was calculated

according to equations (4) and (3); values obtained were substituted in equation (1) and gypsum concentration or solubility determined from the following relationship

$$X = \frac{a}{f} \quad (5)$$

Figure 3 shows the activity coefficient and solubility of gypsum in the presence of sodium chloride, calculated according to the limiting law of Debye-Hückel and equations (4) and (5). Comparing these results with experimental data (figure 2) we see, as might be expected, that the limiting law is not obeyed. We turned next to equation (3) which takes account of the ionic size b ; b was obtained as the sum of crystallographic radii of calcium and sulphate ions, because no data were available on the size of the ions in solution. The calcium-ion radius of 1.06 Å was taken from Goldschmidt's table, and 2.50 Å assumed for the sulphate-ion radius, was based on the structure of sodium sulfate [8]. Thus, a value of 1.06 + 2.05 = 3.11 Å was obtained from the mean effective diameter of the calcium and sulfate ions.

Using the value of 3.11 Å for b in equation (3), the value for the activity coefficient was obtained. This is shown in Figure 4, together with gypsum solubility obtained from equation (5). Again, the curves obtained were much different from experimental curves. We could continue by using more elaborate equations; for instance, those of Hückel [9], Gronvall, La Mer, and Sandved [10]. However, this would lead us to very laborious calculations.

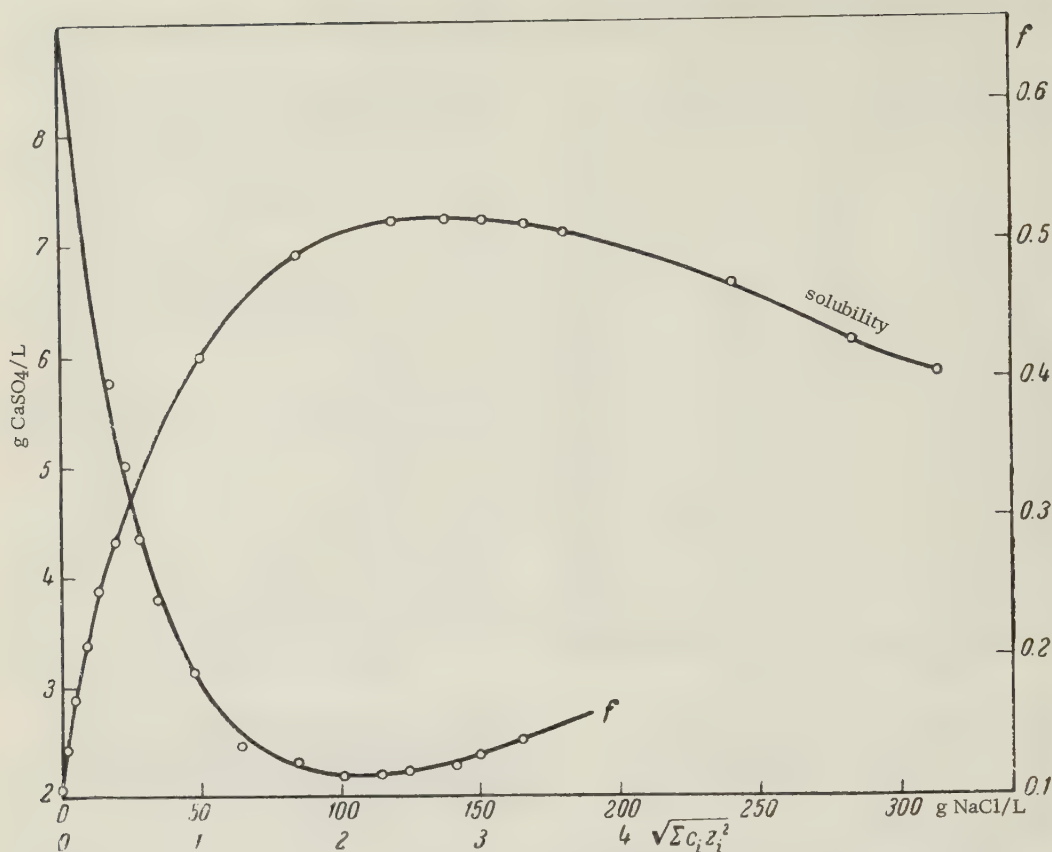


FIGURE 2. Experimental curves of solubility on the coefficient of activity of gypsum in the system $\text{CaSO}_4\text{-NaCl-CO}_2\text{-H}_2\text{O}$ at 25° .

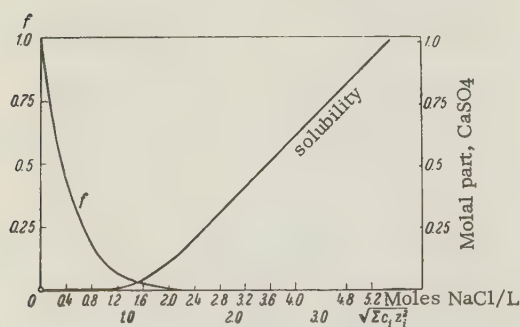


FIGURE 3. Activity coefficient and gypsum solubility in the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$ at 25°C , calculated according to the limiting law of Debye and Hückel.

Therefore, we chose the Debye-Hückel approximate equation (3), as the simplest. Deviations from this equation are unavoidable at rather high concentrations; these result from independent behavior of the ions and should affect the value of b which therefore would change, evidently, as the solution concentration changes.

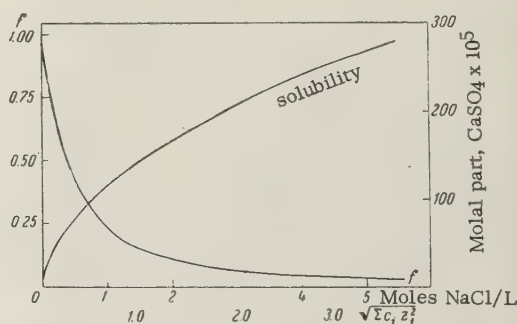
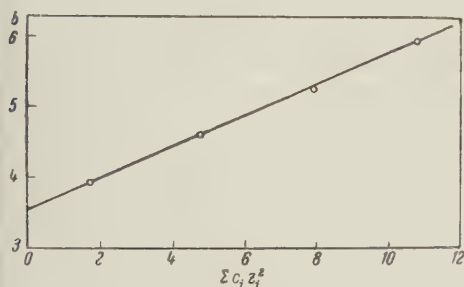


FIGURE 4. Activity coefficient and solubility of gypsum in $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$ at 25°C according to equation (3) with $b = 3.11 \text{ \AA}$.

To find the variation in b as a function of table-salt concentration, the experimental value for the activity coefficient of calcium sulfate and the ionic strength of corresponding sodium-chloride solutions were substituted in equation (3). The change in b as a function of sodium-chloride concentration is given in Table 2 and Figure 5. From these data it can

TABLE 2. Calculated and experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-NaCl-CO}_2\text{-H}_2\text{O}$ at 25° C.

NaCl		Mean effective diameter of ions in solution		Activity coefficient CaSO_4		CaSO_4 molal $\times 10^5$		CaSO_4 g/L	
g/L	mol/L	observed	Calculated according to equation (6)	observed	Calculated according to equations (6) and (3)	observed	Calculated according to equations (6), (3), and (5)	observed	Calculated
19.807	0.339	3.42	(3.70)	0.192	(0.202)	57.2	(54.5)	4.31	(4.12)
35.072	0.600		3.79		0.158		69.6		
50.035	0.856	3.92	3.93	0.138	0.139	79.6	79.1	6.00	6.02
86.103	1.473		4.20	0.121	0.119	91.3	92.4	6.87	7.00
119.867	2.051		4.45	0.115	0.114	96.0	96.5	7.22	7.28
138.756	2.374	4.60	4.60	0.114	0.114	96.4	96.5	7.24	7.27
152.134	2.603		4.70	0.114	0.115	96.3	95.7	7.23	7.21
168.201	2.877		4.82	0.115	0.116	95.6	94.8	7.16	7.14
180.746	3.092		4.91	0.117	0.117	94.3	94.0	7.06	7.08
231.513	3.961	5.28	5.29	0.123	0.123	89.5	89.4	6.65	6.70
283.495	4.850		5.68	0.133	0.133	82.4	82.7	6.10	6.18
317.228	5.427	5.95	5.94	0.140	0.140	78.8	78.6	5.82	5.83

FIGURE 5. Relationship of average effective diameter of ions to the ionic strength of the solution in the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$ at 25°C.

be seen that beginning with about 3 to 4 percent concentration of sodium chloride, the mean effective diameter b of ions in the solutions saturated with gypsum in the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$, can be expressed by a straight line:

$$b = 3.55 + 0.22 \sum c_i z_i^2 \quad (6)$$

Naturally, such an experimental correlation cannot be extended to dilute sodium-chloride solutions in that it does not take into account the ionic strength resulting from calcium sulfate. In dilute solutions this is an important factor compared to the contribution resulting from sodium chloride because calcium sulfate is a 2-2-valent electrolyte.

If one calculates the activity coefficient ac-

cording to equation (3), using numerical values of b found by equation (6), and then finds the solubility by equation (5); the resulting curves would appear similar to those shown in Figure 6. The small circles in the graph indicate experimental points. A comparison of calculated and experimental data is given in Table 2.

This method of determining activity and solubility coefficients by equations (3) and (5), based on assumption of linear increase with concentration of the mean effective diameter of ions in solution was verified in other systems. These systems consist of a saturated solution of gypsum, a sparingly soluble salt, and an added readily soluble salt of varied concentration. These added salts had no common ions with calcium sulfate and, did not form an insoluble complex salt therewith.

Columns 2, 4, and 7 in Table 3 show experimental data on gypsum solubility in aqueous magnesium-chloride solutions at 25° C, obtained by different authors [11, p. 336]. From these data, the ionic strength of solutions was calculated. The ionic strength in dilute solutions (up to about 2 percent magnesium chloride) was calculated from the total concentration of calcium sulfate plus magnesium chloride in solution.

Above 2 percent magnesium-chloride concentration, the concentration of this salt alone was used to calculate the ionic strengths of solutions. Activity of gypsum in its saturated solutions in the presence of magnesium chloride was obtained from experimental data on gypsum solubility in pure water and in dilute magnesium-chloride solutions. As in the preceding system,

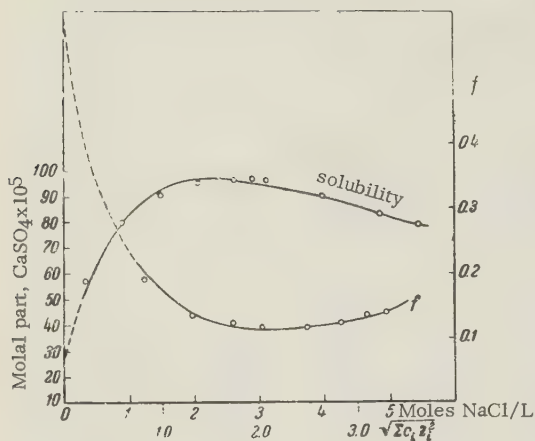


FIGURE 6. Coefficient of activity and solubility of gypsum in the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$ at 25° calculated according to equations (6), (3), and (5). The points designate experimental data.

Experimental values of activity coefficients for two concentrations (points 8 and 10 in table 3) were substituted in equation (3), and then the values of the mean effective diameters b of the ions in solutions of these concentrations, were calculated. In this system it was found that b also changes linearly with the ionic strength of the solution. The linear plot obtained can be expressed by the following equation:

$$b = 3.25 + 0.13 \sum c_i z_i^2. \quad (7)$$

Table 4 and Figure 7 give a comparison of calculated solubility in the system gypsum-magnesium chloride-water, with that obtained experimentally. The calculation was carried out according to equations (7), (3), and (5); mean diameter b of ions in solution for a given magnesium-chloride concentration was found from equation (7); the value for this diameter was substituted in the Debye-Hückel equation (3) to obtain the activity coefficient; and, finally, concentration of solubility of

TABLE 3. Experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-MgCl}_2\text{-H}_2\text{O}$ at 25°C (by different authors [1], p. 336).

Point number	MgCl_2		CaSO_4			d_{\pm}^{ss}	$\sum c_i z_i^2$ MgCl ₂	$\sqrt{\sum c_i z_i^2}$ MgCl ₂	$\sum c_i z_i^2$ total	$\sqrt{\sum c_i z_i^2}$ total	$f - \frac{a}{x}$
	g/L	mol/L	g/L	mol/L	Molal part 10 ⁵						
1	2	3	4	5	6	7	8	9	10	11	12
1	0	0	2.083	0.01530	27.63	0.9993	0		0.1224	0.3499	0.355
2	0.471	0.0049	2.313	0.0170	30.8	0.9974	0.0294		0.1654	0.4067	0.318
3	0.949	0.010	2.506	0.0184	33.3	0.9978	0.060		0.2072	0.4552	0.294
4	1.423	0.015	2.682	0.0197	35.7	0.9982	0.089		0.2470	0.4970	0.275
5	1.896	0.020	2.833	0.0208	37.7	0.9986	0.119		0.2858	0.5346	0.260
6	8.501	0.089	4.255	0.0313	56.5	1.0093	0.536		0.7862	0.8867	0.174
7	19.75	0.207	5.685	0.0418	75.4	1.0194	1.244	1.114	1.5788	1.257	0.130
8	46.64	0.490	7.59	0.0557	100.4	1.0434	2.939	1.715			0.098
9	85.71	0.900	8.60	0.0632	115.5	1.0631	5.400	2.324			0.085
10	121.38	1.275	8.62	0.0693	114.5	1.1022	7.648	2.766			0.086
11	161.89	1.700	8.06	0.0592	107.5	1.123	10.200	3.194			0.091

this is obtained by plotting calcium-sulfate concentration against the square root of total ionic strength of the solution. The straight line so obtained was extrapolated to zero ionic strength; the intercept of the line with the ordinate corresponds to the activity of calcium sulphate, equal to 9.8×10^{-5} molar part. The activity coefficient for gypsum in saturated solutions, in the presence of varying amounts of magnesium chloride, can be obtained easily from the known activity and experimental solubility data by using the equation $f = \frac{a}{x}$, where a is activity, and x is the molal concentration of calcium sulphate (table 3, column 12).

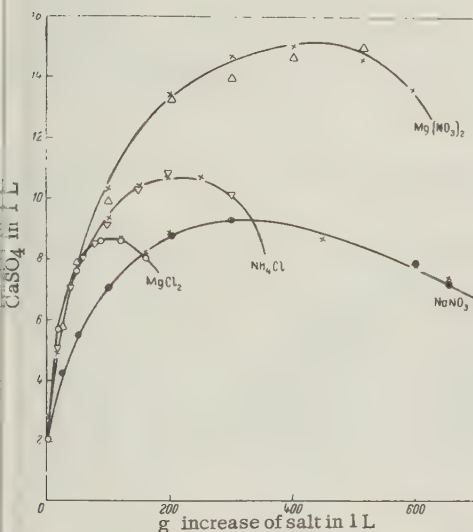
gypsum was determined by equation (5) from the calculated activity coefficient and the known activity.

Table 4 and Figure 7 show that, beginning from approximately 4 percent magnesium-chloride solution, the difference between computed and experimental solubilities does not exceed 1.5 percent of the value itself.

This method of calculation, however, is inapplicable if the magnesium-chloride concentration is too low (see, for instance, point 1 in table 4); because the calcium-sulphate

TABLE 4. A comparison of calculated and experimental data on gypsum solubility at 25° C in the system $\text{CaSO}_4\text{-MgCl}_2\text{-H}_2\text{O}$.

Number of the point plotted	MgCl_2		Mean effective diameter of ions in solution		Activity coefficient CaSO_4		CaSO_4 molal part $\times 10^5$		CaSO_4 g/L	
	g/L	mol/L	observed	Calculated according to equation (7)	observed	Calculated according to equations (7) and (3)	observed	Calculated according to equations (7), (3), and (5),	observed	Calculated
1	19.75	0.207	3.13	(3.41)	0.130	(0.141)	75.4	(69.0)	5.68	(5.23)
2	46.64	0.490	3.62	3.63	0.098	0.098	100.4	100.1	7.59	7.61
3	85.71	0.900	3.95	3.95	0.085	0.086	115.5	114.0	8.60	8.55
4	121.38	1.275	4.25	4.24	0.086	0.086	114.5	114.5	8.62	8.59
5	161.89	1.700	4.58	4.58	0.091	0.090	107.5	108.9	8.06	8.16

FIGURE 7. Solubility at 25° C in the systems: $\text{CaSO}_4\text{-MgCl}_2\text{-H}_2\text{O}$ (circles) $\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$ (black circles) $\text{CaSO}_4\text{-Mg(NO}_3)_2\text{-H}_2\text{O}$ (triangles with vertices up), $\text{CaSO}_4\text{-NH}_4\text{Cl-H}_2\text{O}$ (triangles with vertices down). Calculated values are marked with crosses.

concentration, neglected in our calculations, then becomes considerable compared to the MgCl_2 concentration. Values calculated for points 1 (tables 2, 4, 6, etc.) are, therefore, less accurate and, are placed in parentheses.

This system was not studied when magnesium-chloride concentrations exceeded 16 percent; data on gypsum solubility in such magnesium-chloride solutions, obtained by various authors, are conflicting; the discrepancy might be explained in that analytical determination of calcium and magnesium content is difficult under these conditions.

Our main interest is in the possibility of applying the present computation method to that

part of the solubility curve where it, characteristically, shows a maximum.

Application of the preceding method to the system $\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$ resulted in $a = 9.5 \times 10^{-5}$, and

$$b = 3.16 + 0.10 \sum c_i z_i^2. \quad (8)$$

By substituting these values in equations (3) and (5) a solubility curve is obtained that can be compared to the experimental curve, according to data found in the literature [11, p. 340]. Numerical data for this system are shown in Tables 5 and 6, as well as in Figure 7.

For the system gypsum-magnesium nitrate-water, the activity used for calcium sulfate was the mean of activities obtained for other systems, as the experimental data on gypsum solubility in dilute magnesium-nitrate solutions are insufficient for extrapolation to zero ionic strength.

The mean activity for calcium sulfate is 10×10^{-5} molal.

The activity of a given, sparingly soluble salt in its saturated solution should be constant in the presence of any readily soluble salt.

The result we obtained indicated variations from 11×10^{-5} to 9.5×10^{-5} for the activity of calcium sulfate in different systems; this probably can be explained by experimental errors in the determination of gypsum solubility, especially when magnesium salts were present.

Experimental data of Cameron, Seidell, and Smith [11, p. 336], and the calculations based thereon, are shown in Tables 7 and 8, and Figure 7. The mean effective diameter of ions in solution was calculated from the equation

$$b = 3.20 + 0.025 \sum c_i z_i^2 \quad (9)$$

TABLE 5. Experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$ at 25°C (data of various authors 11, p. 340).

Point number	NaNO_3		CaSO_4			d_{Ca}^{25}	$\sum c_i z_i^2$ NaNO_3	$\sqrt{\sum c_i z_i^2}$ NaNO_3	$\sum c_i z_i^2$ total	$\sqrt{\sum c_i z_i^2}$ total	$f = \frac{a}{x}$
	g/L	mol/L	g/L	mol/L	Molal part 10^5						
1	0	0	2.08	0.0153	27.64	0.9981	0	0	0.1222	0.3496	0.344
2	25	0.294	4.25	0.0312	56.7	1.0163	0.588	0.767	0.8380	0.9154	0.168
3	50	0.588	5.50	0.0404	73.5	1.0340	1.176	1.085	1.4996	1.225	0.129
4	100	1.176	7.10	0.0522	95.5	1.0684	2.353	1.534	2.7700	1.664	0.099
5	200	2.353	8.79	0.0646	120.2	1.1336	4.706	2.169			0.079
6	300	3.529	9.28	0.0682	129.7	1.1916	7.058	2.657			0.073
7	600	7.059	7.89	0.0530	118.1	1.3639	14.117	3.757			0.081
8	655	7.705	7.24	0.0532	110.4	1.3904	15.411	3.926			0.086

TABLE 6. A comparison of calculated and experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$ at 25°C .

Number of the point plotted	NaNO_3		Mean effective diameter of ions in solution		Activity coefficient CaSO_4		CaSO_4 molal part $\times 10^5$		CaSO_4 g/L	
	g/L	mol/L	observed	Calculated according to equation (8)	observed	Calculated according to equations (8) and (3)	observed	Calculated according to equations (8), (3), and (5)	observed	Calculated
1	50	0.588	3.01	(3.28)	0.129	(0.140)	73.5	(67.9)	5.50	(5.10)
2	100	1.176	3.38	3.40	0.099	0.100	95.5	95.0	7.10	7.10
3	200	2.353		3.63	0.079	0.079	120.2	120.3	8.79	8.87
4	300	3.529	3.84	3.87	0.073	0.074	129.7	128.4	9.28	9.27
5	450	5.293		4.22		0.076		125.0		8.72
6	600	7.059	4.52	4.57	0.081	0.082	118.1	115.9	7.89	7.80
7	655	7.705		4.70	0.086	0.085	110.4	111.8	7.24	7.39

TABLE 7. Experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-Mg(NO}_3)_2\text{-H}_2\text{O}$ at 25°C (data of various authors 11, p. 336).

Point number	$\text{Mg(NO}_3)_2$		CaSO_4			d_{Ca}^{25}	$\sum c_i z_i^2$ $\text{Mg(NO}_3)_2$	$\sqrt{\sum c_i z_i^2}$ $\text{Mg(NO}_3)_2$	$\sum c_i z_i^2$ total	$\sqrt{\sum c_i z_i^2}$ total	$f = \frac{a}{x}$
	g/L	mol/L	g/L	mol/L	Molal part 10^5						
1	0	0	2.08	0.0153	27.7	0.9981	0	0	0.1224	0.3499	0.361
2	25	0.169	5.77	0.0424	76.9	1.0205	1.012	1.006	1.3508	1.162	0.130
3	50	0.337	7.88	0.0579	105.5	1.0398	2.023	1.422	2.4858	1.578	0.095
4	100	0.674	9.92	0.0729	133.7	1.0786	4.045	2.011	4.6278	2.152	0.075
5	200	1.348	13.34	0.0980	183.4	1.1498	8.090	2.844			0.055
6	300	2.022	14.00	0.1028	196.4	1.2190	12.134	3.483			0.051
7	400	2.697	14.68	0.1078	211.5	1.2821	16.179	4.022			0.047
8	514	3.465	15.04	0.1105	223.5	1.3553	20.790	4.560			0.045

obtained by the same method used for equation (8). A considerable difference exists between experimental and calculated data in this

system, reaching 2.3 percent for point 4 (table 8). The experimental point probably was not determined accurately, as can be seen

TABLE 8. A comparison of calculated and experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-Mg(NO}_3)_2\text{-H}_2\text{O}$ at 25° C.

Number of the point plotted	$\text{Mg(NO}_3)_2$		Mean effective diameter of ions in solution		Activity coefficient CaSO_4		CaSO_4 molal part $\times 10^5$		CaSO_4 g/L	
	g/L	mol/L	observed	Calculated according to equation (9)	observed	Calculated according to equations (9) and (3)	observed	Calculated according to equations (9), (3), and (5)	observed	Calculated
1	50	0.337	3.03	(3.25)	0.095	(0.103)	105.5	(97.1)	7.88	
2	100	0.674		3.30	0.075	0.073	133.7	137.0	9.92	10.26
3	200	1.348	3.39	3.40	0.055	0.055	183.4	181.8	13.34	13.38
4	300	2.022		3.50	0.051	0.049	196.4	204.1	14.00	14.74
5	400	2.697	3.61	3.60	0.047	0.047	211.5	212.8	14.68	14.97
6	514	3.465	3.65	3.72	0.045	0.0468	223.5	213.7	15.04	14.59
7	595	4.01		3.80		0.0473		241.4		13.61

from its location in Figure 7.

Experimental data of Cameron and Brown [1, p. 337] on gypsum solubility in aqueous ammonium-chloride solutions are given in Tables 9 and 10, and in Figure 7 as well.

mean effective diameter of ions in systems investigated. Values for the first terms of these equations lie between 3.10 and 3.55. It will be recalled that the sum of the mean effective diameter of ions of calcium and sulphate, obtained from crystallographic radii, is equal

TABLE 9. Experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-NH}_4\text{Cl-H}_2\text{O}$ at 25° C (data of various authors [1], p. 337).

Point number	NH_4Cl		CaSO_4			d_{\pm}^{25}	$\sum c_i z_i^2$ NH_4Cl	$\sqrt{\sum c_i z_i^2}$ NH_4Cl	$\sum c_i z_i^2$ total	$\sqrt{\sum c_i z_i^2}$ total	$t = \frac{a}{x}$
	g/L	mol/L	g/L	mol/L	Molal part $\times 10^5$						
1	0	0	2.08	0.0153	27.7	0.9981	0	0	0.1224	0.3499	0.361
2	20	0.374	5.00	0.0367	67.0	1.0042	0.748	0.865	1.0414	1.0204	0.149
3	40	0.748	7.00	0.0514	93.9	1.0103	1.495	1.223	1.9066	1.3808	0.107
4	60	1.122	8.00	0.0588	109.3	1.0158	2.243	1.498	2.7136	1.6473	0.092
5	80	1.495	8.50	0.0624	117.1	1.0208	2.991	1.729			0.085
6	100	1.869	9.10	0.0668	126.4	1.0262	3.739	1.934			0.079
7	150	2.804	10.30	0.0757	146.6	1.0389	5.608	2.368			0.068
8	200	3.739	10.85	0.0797	157.8	1.0518	7.477	2.735			0.063
9	300	5.608	10.10	0.0742	153.8	1.0767	11.216	3.349			0.065

TABLE 10. A comparison of calculated and experimental data on gypsum solubility in the system $\text{CaSO}_4\text{-NH}_4\text{Cl-H}_2\text{O}$ at 25° C.

Number of the point plotted	NH_4Cl		Mean effective diameter of ions in solution		Activity coefficient CaSO_4		CaSO_4 molal part $\times 10^5$		CaSO_4 g/L	
	g/L	mol/L	observed	Calculated according to equation (10)	observed	Calculated according to equations (10) and (3)	observed	Calculated according to equations (10), (3), and (5)	observed	Calculated
1	60	1.122	3.09	(3.26)	0.092	(0.098)	109.3	(102.0)	8.00	(7.54)
2	80	1.495	3.31	3.32	0.085	0.085	117.1	116.3	8.50	8.49
3	100	1.869		3.37	0.079	0.078	126.4	128.2	9.10	9.32
4	150	2.804	3.50	3.50	0.068	0.0685	146.6	146.0	10.30	10.38
5	200	3.739		3.64	0.063	0.065	157.8	153.8	10.85	10.66
6	250	4.673		3.77		0.0637		157.0		10.67
7	300	5.608	3.94	3.91	0.065	0.0641	153.8	156.0	10.10	10.33

The activity for calcium sulphate in this system was taken to be 10×10^{-5} molal on the same basis as that of the preceding system. The following equation was obtained for the mean effective diameter of ions in solution:

$$b = 3.10 + 0.072 \sum c_i z_i^2 \quad (10)$$

Let us compare the equations (6-10) for the

to 3.11 \AA . Because identical ionic dimensions in the crystalline state and in a solution are unlikely, it seems reasonable to assume that the first term represents in fact the value of the mean effective diameter of calcium plus sulfate ions. To understand the possible physical significance of the second term, the following line of reasoning was pursued. All solutions studied represent a combination of

two factors: a saturated solution of gypsum, and a solution of added, readily soluble salt whose concentration varies from zero to saturation. Equations (6-10) consist of two terms: the first pertains to gypsum alone and must remain constant; assuming that the second term pertains to the added salt, then, by analogy, it should express the mean effective ionic diameter of the added salt when its solution is saturated. Let us assume now that a solution is saturated with both gypsum and with table salt. Were our assumption correct, equation (6) for such a solution would appear as

$$b = 3.11 + 2.79,$$

where 2.79 is the sum of the mean crystallographic diameters of sodium and chloride ions. In fact, however, the second term of equation (6) for a solution saturated with sodium chloride is $0.22 \times 10.85 = 2.39$; where the value 10.85 refers to the ionic strength of a saturated sodium-chloride solution.

But 2.79 refers to the saturated sodium-chloride solution. The solution concentration is measured, in such a case, by its ionic strength. For a solution of unit ion strength, the second term should be $\frac{2.79}{10.85} = 0.26$;

yielding 0.26 $\sum c_i z_i^2$ for a solution of ionic strength $\sum c_i z_i^2$.

Equation (6) then assumes the form

$$b = 3.11 + 0.26 \sum c_i z_i^2 \quad (11)$$

Equations (6-10) can be given as

$$b = b_g + \frac{b_{a.s.}}{\sum c_i z_i^2} \sum c_i z_i^2, \quad (12)$$

where b_g is saturated, the mean effective diameter for gypsum (equal to 3.11 Å), $b_{a.s.}$ is the mean effective diameter of the salt added (for example sodium chloride); $\sum c_i z_i^2$ is the ionic strength of a saturated solution of the added salt; and $\sum c_i z_i^2$ is the ionic strength of a given solution of an added, readily soluble salt.

In the system $\text{CaSO}_4\text{-MgCl}_2\text{-H}_2\text{O}$, the mean effective ionic diameter of magnesium chloride can be obtained by taking the arithmetic mean of magnesium and chloride crystallographic diameters:

$$b_{\text{MgCl}_2} = \frac{1.56 + 3.62 + 3.62}{3} = 2.93.$$

Concentration of a saturated magnesium-chloride solution is 4.63 moles per liter, corresponding to an ionic strength $\sum c_i z_i^2 = 27.79$; consequently, the coefficient in the second term of equation (7) becomes $\frac{2.93}{27.79} = 0.11$, and the equation becomes

$$b = 3.11 + 0.11 \sum c_i z_i^2. \quad (13)$$

In our interpretation of the possible physical

significance of constants in the equations for the mean effective diameter of ions, the following factors were considered for systems containing the nitrate ion.

It is known that the positions of nitrogen and oxygen atoms in a nitrate ion can be described as a triangle having N in the center and the three O's at the vertices [12]. The distance $\text{N-O} = 1.21 \text{ Å}$. The N^{+5} radius is very small (0.11 Å), whereas the radius of O can be found from the NO_3^- structure. According to Pauling [12], in the three resonance structures of nitric acid, five out of nine states are those of neutral atoms. Therefore, it is more reasonable to take the oxygen-atom not the ion, dimensions. An oxygen atom, although smaller than the negative oxygen ion, is nevertheless, so large (radius 0.60 Å), that it is possible to describe the NO_3^- as a triangular prism, rather than as a plane triangle. The prism base should be a triangle whose height is twice the distance of N-O , or 2.42 Å. The prism height should be equal to an oxygen-atom diameter (1.20 Å).

Values close to the experimental are obtained when it is assumed that the nitrate-ion prisms are so arranged with respect to other ions in the solution that they make one half the prism height, or the oxygen-atom radius (0.60 Å), equal to the nitrate-ion radius. A mean effective diameter for two spherical ions is obtained by adding their radii, and a mean effective diameter for a spheroidal ion is obtained by addition of one half the prism height and the spherical-ion radius.

Thus, the mean effective diameter for sodium nitrate in the system $\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$ is $b = 0.98 + 0.60 = 1.58 \text{ Å}$.

A saturated sodium-nitrate solution contains 7.71 moles per liter, which corresponds to an ionic strength $\sum c_i z_i^2 = 15.42$. The coefficient of the second term in equation (8) becomes $1.58/15.42 = 0.10$; and equation (8) for this system is

$$b = 3.11 + 0.10 \sum c_i z_i^2. \quad (14)$$

The mean effective diameter for magnesium-nitrate ions in the system $\text{CaSO}_4\text{-Mg(NO}_3)_2\text{-H}_2\text{O}$, can be obtained as the arithmetic mean of the effective diameters of one magnesium ion and two nitrate ions:

$$b = \frac{1.56 + 1.20 + 1.20}{3} = 1.32. \quad \text{A saturated}$$

magnesium-nitrate solution contains 4.01 moles per liter; or, $\sum c_i z_i^2$ of the saturated solution is equal to 24.06.

In such a case

$$\frac{b_{a.s.}}{\sum c_i z_i^2} = \frac{b_{\text{Mg(NO}_3)_2}}{\sum c_i z_i^2} = \frac{1.32}{24.06} = 0.05 \quad (15)$$

and $b = 3.11 + 0.05 \sum c_i z_i^2$.

It was impossible to explain the magnitude of 0.7 in equation (10) of the system $\text{CaSO}_4\text{-H}_2\text{O}$ on the basis of crystallographic dimensions for the ammonium ion. The volume of an ammonium ion in solution proves to be considerably smaller than in the crystalline state. It is possible that this discrepancy could be explained by hydrogen-bond interaction between NH_4 and H_2O ; this may lead to strong polarization of the ammonium ion and to decrease in its volume (if the ion rotates).

Comparison of experimental equations (6-10) with equations (11-15) (table 11) in which empirical constants are replaced by those obtained from crystallographic data, indicates that the crystallographic dimensions of ions

in the system. In such cases, equilibrium sets in very slowly; investigation becomes much easier when it is possible to find the solubility by calculation.

CONCLUSIONS

1. For the system $\text{CaSO}_4\text{-CO}_2\text{-H}_2\text{O}$ containing sodium chloride in amounts ranging from zero to saturation, solubility is investigated experimentally at 25° C. The curve obtained has a maximum.

2. Solubility of gypsum in aqueous sodium-chloride solutions (up to saturation) is calculated according to the Debye-Hückel equation; it is

TABLE 11. Mean effective diameter of ions in the systems studied

Systems	Experimental equations	Equations obtained by using crystallographic values for the empirical constants in equations (6-9)
$\text{CaSO}_4\text{-NaCl-H}_2\text{O}$	$b = 3.55 + 0.22 \sum c_i z_i^2$ (6)	$b = 3.11 + 0.26 \sum c_i z_i^2$ (12)
$\text{CaSO}_4\text{-MgCl}_2\text{-H}_2\text{O}$	$b = 3.25 + 0.13 \sum c_i z_i^2$ (7)	$b = 3.11 + 0.11 \sum c_i z_i^2$ (13)
$\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$	$b = 3.16 + 0.10 \sum c_i z_i^2$ (8)	$b = 3.11 + 0.10 \sum c_i z_i^2$ (14)
$\text{CaSO}_4\text{-Mg(NO}_3)_2\text{-H}_2\text{O}$	$b = 3.20 + 0.03 \sum c_i z_i^2$ (9)	$b = 3.11 + 0.05 \sum c_i z_i^2$ (15)
$\text{CaSO}_4\text{-NH}_4\text{Cl-H}_2\text{O}$	$b = 3.10 + 0.07 \sum c_i z_i^2$ (10)	

The general form of equations (12-15) is:
$$b = b_g + \frac{b_{a.s.}}{\sum_{\text{sat.}} c_i z_i^2} \sum c_i z_i^2, \quad (11)$$

where b_g is the mean crystallographic diameter of calcium and sulfate ions, equal to 3.11 Å,

$b_{a.s.}$ is the mean crystallographic diameter of an added salt,

$\sum_{\text{sat.}} c_i z_i^2$ is the ionic strength of a saturated solution of table salt, and

$\sum c_i z_i^2$ is the ionic strength of a given solution of the added salt.

The solubility of the added salt, are the basis of the first attempt to explain the complex phenomenon of dependence of the mean effective diameter of ions upon concentration of the systems composed of gypsum and aqueous solutions of readily soluble salts that have neither ammonium ions, nor form chemical compounds with gypsum.

The results of this study show that the Debye-Hückel equations facilitate investigation of the solubility of a poorly soluble salt when readily soluble salts are present. This is true especially in our system when one more component such as calcite (left undiscussed here) participates

assumed that the mean effective diameter of ions in solution is a function of concentration.

3. A linear equation is obtained for the relationship between effective ion diameters in the system $\text{CaSO}_4\text{-NaCl-H}_2\text{O}$ and sodium-chloride content in the solution.

4. The calculation method to determine gypsum solubility is verified on the systems $\text{CaSO}_4\text{-MgCl}_2\text{-H}_2\text{O}$, $\text{CaSO}_4\text{-NaNO}_3\text{-H}_2\text{O}$, $\text{CaSO}_4\text{-Mg(NO}_3)_2\text{-H}_2\text{O}$, and $\text{CaSO}_4\text{-NH}_4\text{Cl-H}_2\text{O}$ according to experimental data of various authors.

5. It is shown that the experimental linear equations that correlate effective ion diameters with concentrations, can be presented in the following approximate form:

$$b = b_g + \frac{b_{a.s.}}{\sum_{\text{sat.}} c_i z_i^2} \sum c_i z_i^2, \quad (11)$$

where b is the mean effective diameter for ions in solution; b_g , the mean crystallographic diameter for calcium and sulfate ions, is equal to 3.11 Å; $b_{a.s.}$ is the mean crystallographic diameter for added salt; $\sum_{\text{sat.}} c_i z_i^2$ is the ionic strength of a saturated added-salt solution; and $\sum c_i z_i^2$ is the ionic strength of a given added-salt solution.

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USE OF REFLECTION-SEISMIC METHODS IN THE EXPLORATION OF DEEP BEDS¹

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• translated by T. Gretener •

ABSTRACT

Observation of reflections reported from very great depths, from reflection-seismic measurement by Seismos GmbH. [Ltd.] (that were commissioned by the Preussischen Bergwerks- und Hütten AG. [Inc.], Mobiloil AG. in Germany, and by Wintershall AG. in various project districts of Germany) will be discussed. The most salient groups, most probably, could be related to the Conrad and Mohorovičić discontinuities. -- Auth.

G. Schultz [12] was the first to report on deep reflections in the course of routine reflection-seismic research. More than 230 reflections with anomalously long travel times, in the order of 5 sec, have been observed in the Nahe-Senke area. In addition, at some places, strong reflections of 4 sec travel time were observed. Schultz uses a velocity of 500 m/sec [18,040 ft/sec] for the deeper sediments plus granite in his depth calculations. For the upper beds, the known local velocities were used. By these assumptions, the depth for the 4 sec reflection group comes to 10.4 km [6.5 mi]. Using a Conrad 6.5 km/sec velocity of 6000 m/sec [19,680 ft./sec] for the rocks beneath this discontinuity, the 5 sec reflections must be placed at 13.4 km [8.3 mi]. The latter reflections are believed to originate from the Conrad discontinuity. The 4 sec reflection might belong to a granite-gneiss discontinuity as assumed by Förtsch. A statistical analysis of the reflections indicates that the lower discontinuity dips about 5° south-southeast.

It is remarkable that these deep reflection groups were found also in the Rastatt-Achern area in the Rhein graben [1]. On a N-S profile the reflections occur at 7.5 to 8 sec travel time with no apparent dip. On a W-E profile N of Bühl, the deep horizons dip steeply towards the west.

These deep reflections cannot be interpreted as multiples because they cannot be separated from higher horizons. This holds for the Nahe-Senke area as well. There is also a strong lack of energy on seismograms between the last reflections for the sedimentary sequence

and the strong deep reflections.

Interpretation of the strong dip toward the east on the W-E section is hampered by the fact that the shotpoints were located very close to the Black Forest. Dip and location of the reflecting plane are, therefore, certainly falsified by "refraction influences".

An interpretation of the profile with reasonable assumptions of velocity distribution at the fringe of the Black Forest reduces the dip toward the east; but, even rather extreme assumptions cannot nullify the dip completely.

It might be possible to reduce even the "eastward dip" by making some very special assumptions. However, use of rather reasonable assumptions shows that the dip is still present. This leaves the question open: Is this dip real or apparent? Comparison of depth of the discontinuity with depth in neighboring areas, indicates that at least part of the dip is real.

The depth in the Rastatt-Achern area is approximately 17 to 18 km (10.6 to 11.2 mi). Comparison of depth and character of the reflections indicates, very probably, that they originate from the same discontinuity as those in the Nahe-Senke area; i. e. from the Conrad discontinuity.

In the event that deep horizons cannot be plotted as continuous profiles, statistical analysis still allows reasonable interpretation. In these cases, frequency curves for deep reflections were plotted. On these curves is given the total number of reflections observed for the true interval (0.2 sec) in the area. The maxima of these curves represent either multiple reflections or true reflection horizons. The decision as to which of the two is valid has to be made with the seismogram.

¹Translated from *Ein Beitrag der Reflexionsseismik zur Erforschung des tieferen Untergrundes*; Geologische Rundschau, v. 46, no. 1, 1957, p. 17-26.

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The method has been applied in the Nahe-Senke area with success. The dip of the Conrad discontinuity was found to be 5° S, in good

agreement with Schulz's results.

The regional picture of deep reflection horizons can be gained by comparing frequency curves for different areas and correlating characteristic maxima.

Observations in the Lippstadt and Murnau areas give additional information on regional behavior of the Conrad discontinuity.

Near the town of Lippstadt, unusually strong reflections are observed between 5.5 and 6.5 sec; usually there are several strong

reflections in that time interval: a band of reflections with strong energy. It seems, therefore, that there are a series of discontinuities rather than one sharp discontinuity. The assumption of a boundary zone rather than a sharp boundary seems justified.

Figure 1 gives part of three continuous seismograms taken near Lippstadt. Several strong reflections occur between 5.5 and 6.5 sec. On some seismograms of this area, there are good reflections at even greater depth. They cannot be interpreted at the present time.

Figure 2 gives the frequency curve of deep

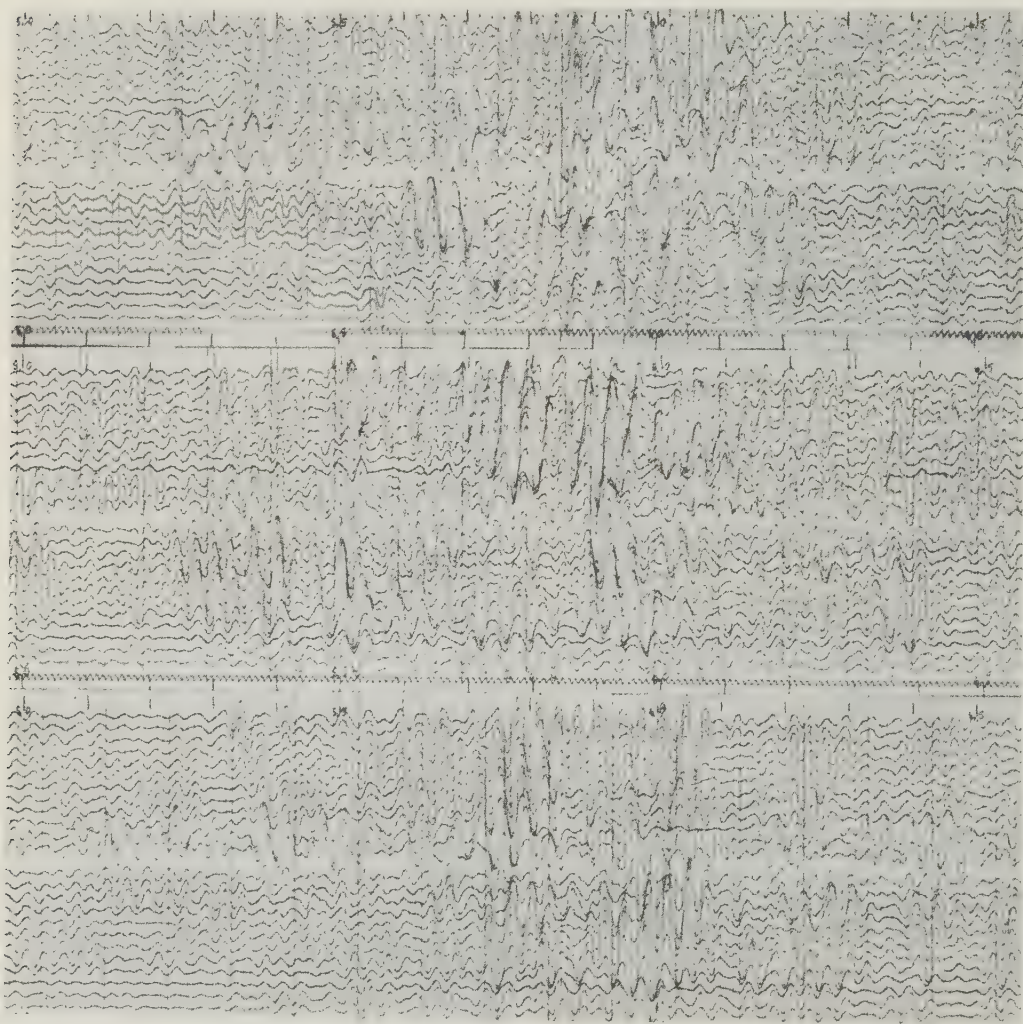


FIGURE 1. Section from three seismograms of Seismos GmbH. from the Lippstadt project area (Contractor Mobiloil-AG. in Germany). The portion between 5.0 and 6.5 sec reflection time is shown. One can detect much stronger reflections, than would probably originate from the Conrad discontinuity range.

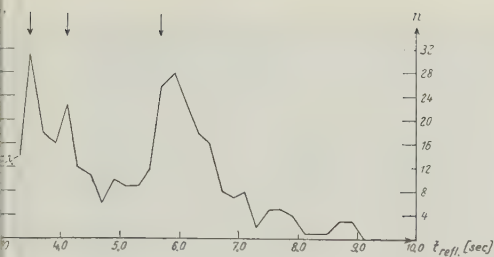


FIGURE 2. Frequency distribution of deep reflection in the Lippstadt project area. The observed quantity is depicted by deep reflections as a function of travel time with interval breadths of 0.2 sec.

reflections for the above area. A wide maximum between 5.5 and 6.5 sec is indicated clearly. Most probably, it has to be correlated with the Conrad discontinuity. Well defined maxima are present at 3.5 and 4.0 sec. These will be explained later.

For the depth determination, the following velocity distribution is assumed for the Lippstadt area. To 0.5 sec, 2800 m/sec [9,184 ft/sec] to 0.7 sec, 4000 m/sec [13,120 ft/sec] to 4.0 sec, 5000 m/sec [16,400 ft/sec] to 6.0 sec, 5500 m/sec [18,040 ft/sec]. Using the above velocity distribution, one finds that the Conrad discontinuity comes to 13.5 to 14 km [8.4 to 8.7 mi]. The upper maxima lie at 5 km [5 mi] and 9.5 km [6 mi]. The data are sufficient to enable one to decide whether or not the small maximum of 8.8 sec is meaningful. Assuming a velocity of 6000 m/sec [19,440 ft/sec] for the work underneath the Conrad discontinuity, the latter maximum would correspond to a depth of 23 km [14.3 mi].

Interpretation of these reflections as multiples is highly improbable because the upper beds are strongly folded and thrust. Furthermore, quality of the upper reflections is rather poor compared to those at depth.

The same is true for the Murnau area [1] in the folded Molasse region of Upper Bavaria. Because of absorption, refraction, and scattering, multiple reflections are highly improbable.

Figure 3 shows the frequency curve for deep reflections in the Murnau area. In addition to the maximum at 6.5 sec, a second is visible at about 11 sec. The 11 sec reflections are well developed in areas of good energy radiation; they often correlate over several records. Earlier and later reflection groups are common; it is quite probable that here, too, there is a group of discontinuities rather than one sharp boundary. In this area shots of 50 to 75 kg [10 to 165 lb] were used.

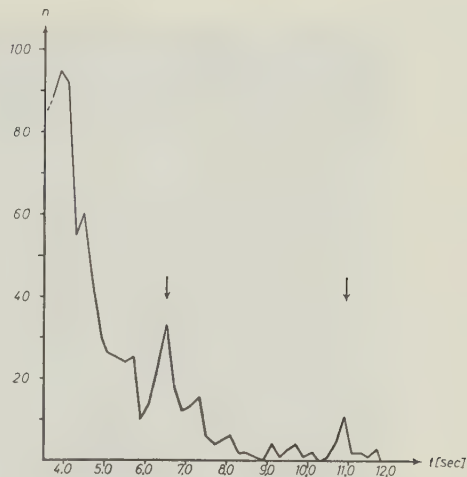


FIGURE 3. Frequency distribution of deep reflections in the Murnau project area.

Figure 4 shows part of 3 records with reflection groups of 11 sec. They are preceded by a long, quiet area, whereas the upper horizons clearly are dipping.

Assuming an average velocity of 4500 m/sec [14,580 m/sec] for the folded Molasse down to a reflection time of 3.0 sec, 5500 m/sec [18,040 m/sec] to 4.5 sec, and 6000 m/sec [19,680 m/sec] for the deeper horizons the following depths are calculated.

4.4 sec	10.5 - 11.0 km
6.5 sec	17.0 - 18.0 km
11.0 sec	30.0 - 31.0 km

More accurate depth determinations are not possible at the present because velocity distribution in the folded Molasse of upper Bavaria still is insufficiently known. It is estimated that errors in the above depth determination do not exceed 5 percent.

It seems reasonable to identify the 11 sec reflections with the Mohorovičić discontinuity registrations of local earth quakes and blasts that indicate a depth [5, 7, 8, 9] of approximately 30 km [18.6 mi] for that discontinuity near the Alps. The 6.5 sec reflections most probably originate at the Conrad discontinuity. This group, however, varies quite a bit and often is missing even when energy conditions are favourable. Sometimes the 6.5 sec reflections are replaced by others that occur at 7.0 sec.

They, too, appear to belong to the Conrad discontinuity, thus indicating greater depth for that reflecting plane.

In some places, reflections were observed at 3.5 sec; this can be correlated over several

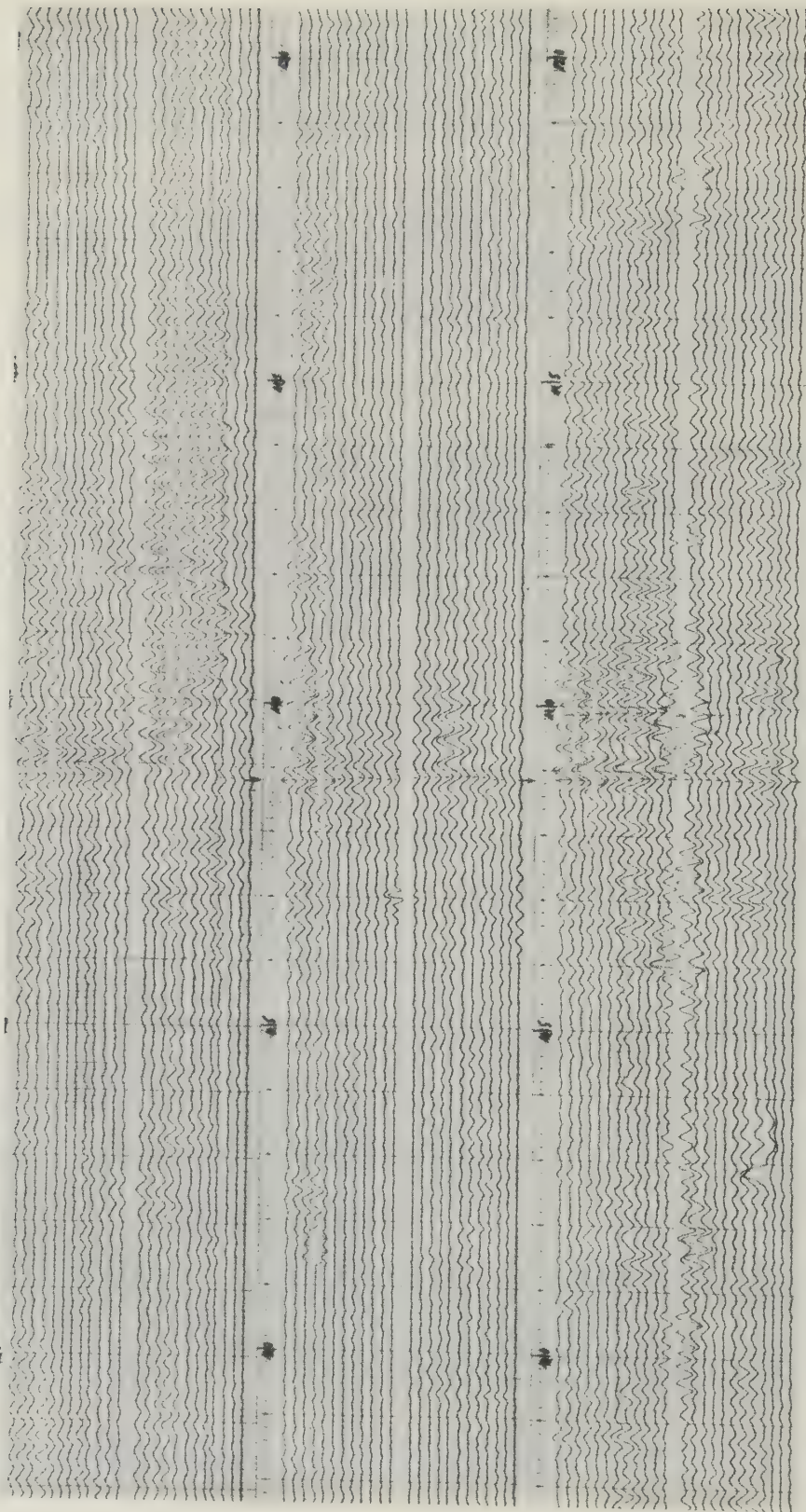


FIGURE 4. Section from three seismograms of Seismos GmbH, from the project area Murnau (Contracting Company Florentine I) with very deep reflections, at approximately 11 sec reflection time.

records. They originate at a depth of 8 to 8.5 km [5 to 5.3 mi] and might be reflected from the top of the basement.

In some places, very strong-reflections occur at 4.5 sec. These might originate at the Försch discontinuity (granite-dionite) but more data are needed before this question can be decided. Comparison of the frequency curves for Murnau and Lippstadt suggests that the 3.5 to 4.0 sec reflections in Lippstadt also might originate at the Försch discontinuity. Schulz, too, attached some reflections in the Nahe-Senke area to this hypothetical discontinuity.

Deep reflections were observed for the first time in northern Germany in the Vorhop area. The reflections occur at 5.5 to 6 sec. Making reasonable assumptions on the velocity distribution, the depth is 12 to 15 km [7.5 to 9.3 mi]. Again the reflections seem to originate in a rather wide zone. These reflections are not as good as those mentioned before, but they do fit well with the results of the Helgoland blast.

There follows a table showing deep reflections recorded by Seismos in the course of normal reflection work.

Region	Försch discontinuity	Conrad discontinuity	Mohorovičić discontinuity
Vorhop	-	12(-15)(km)	-
Lippstadt	8 - 9.5 (km)	13.5-14 (km)	-
Nahe-Senke	10.5 (km)	13.5 (km)	-
Rastatt	-	17 -18 (km)	-
Murnau	10.5-11 (km)	17 -18 (km)	30 - 31 (km)

The present depth determinations fit well with data gained from earth quakes and blasts; and depth of the Conrad discontinuity, in different areas confirms the southerly dip of this reflecting plane. Also it strengthens the possibility that this discontinuity is continent wide

[2, 11]. Certain irregularities in this discontinuity are indicated. In the Rastatt area, the depth is anomalously high; and, it is possible that the Rheingraben is reflected in the Conrad discontinuity. The sections near Bühl support this interpretation.

Assuming the Försch discontinuity to be real, a slight southerly dip of this discontinuity is indicated.

Junger [6] has reported some deep reflections from North America. These reflections from approximately 18 to 21 km depth have been attributed to the lower face of the granite zone, i. e. The Conrad discontinuity. The Mohorovičić discontinuity has been determined by several blasts in the U. S. and U. S. S. R. [9].

The present paper shows that there is a possibility of obtaining deep reflections with normal reflection instruments. The combination of these observations might give valuable information on structure of the crust.

At this time, I should like to express my gratitude to the Preussischen Bergwerks- und Hütten AG., Mobiloil AG. in Germany, the

Gewerkschaft Elwerath and Wintershall AG. for their kind assistance, as well as for the permission to publish the materials reproduced here. Similarly, I should like to thank Seismos GbmH. for their approval for publication of this work.

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- [This translation is published with the author's permission through the cooperation of the Shell Oil Company of Canada for whom the translation was originally made. The attention of researchers desiring later information on this program of seismograph surveying by recording reflections with long travel time is drawn to Dr. Dohr's paper, Über die Beobachtungen von Reflexionen aus dem tieferen Untergrunde in Rahmen routinemässiger reflexionsseismischer Messungen: Zeitschrift für Geophysik, v. 25, no. 6, 1959, p. 280-300. --M. R.].

FLUCTUATIONS OF THE ARAL SEA LEVEL¹

by A. S. Kes²

• prepared by the U.S. Joint Publications Research Service •

ABSTRACT

The Aral sea (located approximately 45°N-60°E), outlet basin of the Amu-Darya and Syr-Darya rivers, is similar to other basins enclosed by a desert area in that its level is very unstable. In the past, these rivers have been inconstant in the amount and direction of their flow; and, thus, have played a considerable role in the sea-level fluctuation. Evidence of former, higher sea levels of the Aral is indicated by its terraces, composed of sea sedimentation. The highest terrace is composed of gray sand, and shells of the mollusk *Cardium edule*. Below this, there occurs a second terrace of light blue-gray carbonate loam; lower terraces of solonchak (salt marsh) extend to the Aral shoreline. Maximum elevation of past sea-levels, according to terrace elevations, was approximately 49 to 57 meters. Occurrence of freshwater mollusk shells of *Anodonta* with those of *Cardium edule* found in situ in alluvial deposits along the downstream channels of the Zhana-Darya river, indicate the frequent changes in regime of the Aral sea. The maximum level of its transgression probably occurred, in the northern Aral region, about 3,000 B. C.; this was based on distribution of *Cardium edule* and, of sites of the primitive Celtaminarian peoples. These same sites in the southeastern Aral region indicate the maximum to have occurred from 700 to 1,000 B. C. In recent years, the highest level recorded for the Aral was 53 meters in 1956; the lowest, 51.5 meters, in 1920. The sea-level fluctuation curve for the Aral sea is in opposition to that of the Caspian, but in agreement with that of Lake Balkhash. Opposition of the Aral and Caspian fluctuation curves probably results from intensification of glacier melting during periods of drought, thus swelling the Amu-Darya and Syr-Darya rivers and, consequently, raising the Aral sea level. Aral sea-level fluctuations, then, result from drastic changes in climate of long duration throughout considerable areal extent and, depend on factors causative of glacier ablation where meltwaters drain into rivers that feed the Aral sea.--D. D. Fisher.

The Aral sea, terminal basin of central Asia's largest rivers (the Amu-Darya and the Syr-Darya), has a very unstable level, similar to other and-locked basins of a desert region.

The condition of this vast sea-lake depends basically on influx of river waters; that is why the Aral sea became a vast basin only after the Amu-Darya and Syr-Darya rivers began to empty entirely into the wide Aral basin. Both of these rivers possess great inconstancy: during the latest stages of geological history they have shifted their course repeatedly, at times discharging into the Aral sea, at times turning aside, and partially or completely emptying their waters into other basins of central Asia. When the Amu-Darya diverted its course for a long or short period of time, e.g., toward Sarykamys, the Aral sea's level dropped sharply. However, even at times

when both the Amu-Darya and Syr-Darya discharged completely into the Aral sea, the level of the Aral still fluctuated, though to a lesser degree. These fluctuations, very frequently witnessed by human beings, left very clear traces behind them.

Fluctuations of the Aral sea level have long attracted the attention of scientists. True, the first investigators talked mostly about terraces, evidences of a former higher sea-level, and made conclusions about the occurrence of desiccation of the Aral sea. However, evidence also exists of lower level of the Aral, i.e. the washed out riverbeds on its sea-floor; these can be traced clearly among the numerous islands of the Akpektin archipelago in the southeastern part of the sea-lake (fig. 1). Here, general contours of the river beds with the adjoining lowlands, extensions of dry beds of the Zhana-Darya (ancient deltaic tributary of the Syr-Darya) and the Akcha-Darya (ancient deltaic tributary of the Amu-Darya) as well as certain segments of meandering, deeper areas of the river beds, are visible through the water in the shallowest areas of the gulfs.

Translated from *O kolebaniyakh uroviya Aral'skogo Morya*: Priroda, no. 1, 1958, p. 95-99. JPRS (NY)L-342.

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The Aral sea terraces were formed by sea sedimentation consisting largely of mollusk shells that, in places, almost completely cover

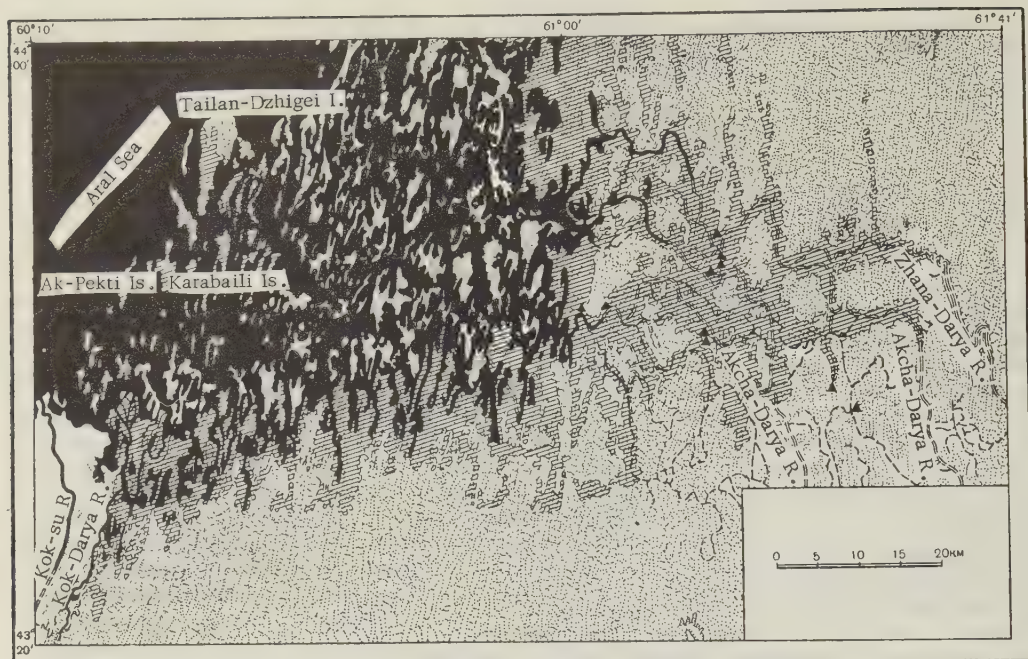








FIGURE 1. Map of southeastern littoral zone of the Aral sea.

- | | |
|---|---|
|  1-sea water |  4-river beds flooded by the sea |
|  2-regions previously flooded by the Aral sea |  5-dry river beds |
|  3-desert area |  6-stations of primitive man |

the surface of the once-submerged plain. The highest terrace, corresponding to the maximal transgression, is formed of gray sands and *Cardium edule*. These sands are mostly airborne; the terrace has an ashen relief of meridially extended ridges up to 3 to 5 m in height. The maximum width of the terrace is 1 to 1.5 km. It stretches along dry beds of the Zhana-Darya and Akcha-Darya and, in irregular tongue-shaped forms, embeds itself along depressions among high ridges in the region of the ancient ashen relief in sands of the southeastern Aral sea area. However, this terrace is hardly ever so clearly marked: More frequently, traces of maximum stagnation of the Aral are marked solely by a dark bluish-gray streak traced on a definite level against a background of yellow slopes of high sand ridges; beneath these, sea-shells are always encountered.

At 1.5 to 2 m below this horizon there is a sea terrace formed of light bluish-gray carbonate loam saturated with shells of sea mollusks of the same type as those of the upper terrace. Small areas of this terrace are encountered principally along river beds and in the lowest basins and depressions. Extending still lower are the solonchak areas that con-

tinue as far as the surface of the Aral sea water.

All of these three surfaces, at the present time located above sea-level, serve as evidence of previously higher water levels of the Aral sea. Whenever the sea level rose, the sea inundated considerable expanses of the lower shores and ingressed deeply the inter-ridge depressions. It moved particularly high up along the river bed of the Zhana-Darya; *Cardium edule*, *Hydrobia*, *Caspia* and other sea shells are found as far as the latitude of the Dzhaman-Bay Kolkhoz, 100 km above the present southeastern bays of the Aral sea.

What, then, was the maximum height of the Aral sea level?

As recently as the turn of our century L. S. Berg wrote that the level of a terrace containing *Cardium edule* was situated 4 m above the sea level of the present Aral, i.e. on the absolute mark of 54 m [1]. However, A. L. Yanshin has demonstrated graphically that in the northern Aral region, the terrace containing *Cardium edule* occurred at various levels, from the 53 to the 64 m mark; that the lower levels were coordinated [primochony] with

cient synclines, and the higher levels, with ancient anticlines. This led A. L. Yanshin to the conclusion that sediments of the ancient transgression were dislocated [2]. Consequently, from data obtained from the northern Aral region it is impossible to determine the actual level of maximum transgression of the Aral sea.

The southeastern region of the Aral sea is situated in the area of the vast, ancient Syrdarian tectonic depression. Leveling has shown that maximum height of the upper terrace with *Cardium edule* is here 3.5 to 4 m higher than the present water level in the Aral sea (fig. 2). The absolute mark of the mean

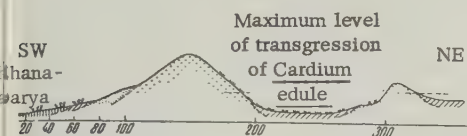


FIGURE 2. Profile of the right bank of the Zhana-Darya river, 16 km east of the eastern gulfs of the Aral.

- 1-yellow sands
- 2-gray sand
- 3-marl clay loam
- 4-heavy [fine?] solonchak clay loam
- 5-Sarsazan [transliteration] (*Halocnemum strobilaceum*)
- 6-Tortula desertorum
- 7-Station Aksaga-2

level of the present Aral sea is 53 m. Consequently, absolute altitude of the highest ancient transgression on the southeastern shore is 56 to 57 m; i.e. it is considerably higher than that indicated by L. S. Berg. If we take into account that the lowest Aral sea level was recorded at the absolute mark of 49.25 m (1880) [3], amplitude of its sea level fluctuation for the recent historical period is equal to approximately 8 m.

Most investigators believed Aral sea-level fluctuations to be related to climatic changes. According to B. D. Zaykov, for a long time the Aral received more than 90 percent of the water from surface inflow; 73.4 percent of this is contributed by the Amu-Darya, and 23.2 percent, by the Syr-Darya. Almost without exception, however, the water is lost through evaporation.

Consequently, the condition of the Aral sea

depends, on the one hand, on climatic changes occurring directly in the Aral region, and, on the other hand, on changes in flow of the Amu-Darya and Syr-Darya; this condition also is related basically to climatic changes in the mountains.

Could such sharp fluctuations in sea level, reaching 8 m, have occurred without considerable climatic fluctuation? To answer this question, it would be interesting to study fluctuations of the sea level over a period of time when meteorological and hydrometrical observations of the Aral sea were being conducted. The findings show that during this period of time the Aral sea level fluctuated sharply; from 1880 to 1955 the sea level rose at least 4 m. Moreover, it is a known fact that no catastrophic changes in the climate were recorded during this period. It is possible that during the period of maximal transgression the Aral received some additional water supply from, for example, the Chu and Sary-Su rivers, and perhaps from the Turgay river as well. However, judging from the fluctuations taking place in front of our very eyes, we may assume that such a rise in sea level can occur even without severe and long-lasting climatic changes.

The Aral sea always has been distinguished by the marked instability of its level; inundation of its shores never has been of long duration. This is evidenced by absence in this region of any significant sea sedimentation; as a rule, this sedimentation is marked only by layers measuring a few centimeters. The *C. edule* *Hydrobia* and other shells are encountered everywhere, but only on surfaces and in locations of thin surface deposits (fig. 3).

For scores of kilometers in the down-stream channels of the Zhana-Darya river, *Cardium edule* shells and shells of the fresh water mollusk *Anodonta* can be found among the alluvial deposits. Despite its great fragility, this shell usually is intact with both halves preserved; *Anodonta*, therefore, is not carried inertly by the river stream into the sea gulfs but is located in situ. The joint occurrence of both fresh and sea-water shells can be explained only by the frequent changes of the Aral sea regime.

On the basis of the distribution of stations [stoyanki] of Celtaminarian Culture along the shores extant during the transgression east of the Aral gulf, A. L. Yanshin reaches the conclusion that the transgression, which deposited sands containing *Cardium edule* on the Aral shores, occurred in the third millennium B. C.

Working jointly with the Khorezm expedition, we became convinced that the primeval stations were indeed coordinated in time with Aral shores of maximum transgression. Most of the stations encountered were situated on slopes of sand ridges in the immediate vicinity of

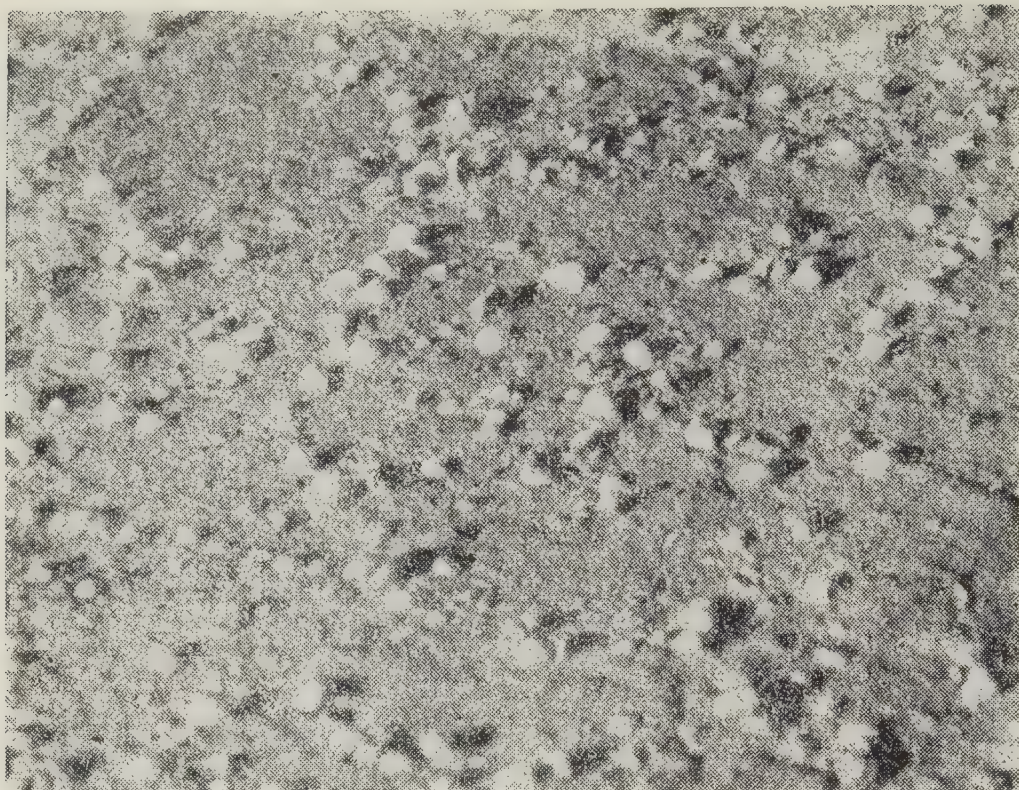


FIGURE 3. Shells of mollusks Cardium edule covering the surface of the terrace of maximum transgression of the Aral sea.

former coastlines or closest to submerged basin areas, yet always near the river beds. We gained the impression that people had lived on the shores of the sea gulf into which the tributaries [pretoki] of the Zhana-Darya emptied their waters.

However, according to S. P. Tolstov's estimation, stations on the southeastern shores date back to the seventh to tenth Centuries B.C.; thus, they are almost two thousand years younger than stations of the northern Aral region. Therefore, it may be assumed that the maximum ancient Aral transgression had occurred in the northern and southeastern Aral regions at different times; this, of course, would be impossible. It appears to us that this can be explained by the assumption that the Aral reached such a high level repeatedly and at different times.

Today, we can speak of at least two high stands of the Aral, which attained the closest maximum level in the third, and at the beginning of the first, millennium B. C. In the intervals between those transgressions and later, the Aral sea-level dropped sharply, a fact evidenced by river beds submerged at present by the sea and, by lower terraces and traces

of shore lines.

The condition of the Aral sea in the 18th, 19th, and at the beginning of the 20th Century has been described rather fully by L. S. Berg [1, 3]; and, for subsequent years (up to 1940), by B. D. Zaykov [4].

Behavior of the Aral sea in recent years can be judged by observations carried out on a number of stations situated on shores and islands of the Aral sea. Preliminary acquaintance with the observational data makes it possible to establish that after the level of the Aral sea had risen 3 m from 1880 to 1910, and, while continuing to remain on a high level it changed, nevertheless, rather drastically. The Aral sea reached the highest levels in 1915, 1925, 1935, 1936, 1945, and 1955-1956, remaining at marks close to 52 m and in 1956, even above 53 m. The lowest level during these years was reached in 1920 when the water level line dropped to 51.5 m. The amplitude between the lowest level of 1920 and highest of 1955 amounted to almost 2 m.

The sharp rise in water level of the Aral sea-level line in recent years, caused new, vast inundations of the southeastern Aral region. At



FIGURE 4. Gulfs of the Aral sea, situated on beds between depressions.

At the present time, this flooding still continues to occur very intensively (fig. 4). Comparison of the limits of water expansion in the river beds, recorded by aerial photography in 1952, with that observed in October 1956, shows that for the past four years, water along the Zhana-Darya river bed moved 22 km upward and 10 km upward along one of the southern beds of the Akcha-Darya river.

Last year's tracks of an automobile passing over the dry bottom of the river bed now are under water. At a distance of 0.5 m upstream from the limit of the water expansion, the bottom of the Zhana-Darya river bed proved to be so swampy that it was impossible to traverse. Caravan trails, over which camels passed not long ago, also are submerged in the lowest places.

The loamy expanses surrounding the gulfs are saturated with water and have been formed into swampy, moist solonchaks (salt marshes). All of this attests that the present sea level is higher than that of previous years.

In conclusion, it is interesting to note once again that the curve of Aral sea-level fluctuation is in opposition to the fluctuation curve for the Caspian sea level. In years of drought, when the melting of glaciers intensified, the current of the Amu-Darya, and Syr-Darya rivers increased, and the water level of the Aral rose; while in the Caspian, however, it fell, and

vice versa. On the other hand, the course of fluctuation of the level of the Balkhash (located in similar geographic environments) is similar to fluctuations of the Aral. Causes of the Aral sea-level fluctuations are the result not only of severe and long-lasting climatic changes throughout vast areas but, of transitory fluctuations that cause intensification of snow melting in the mountains and, of the discharge of the greatest rivers in central Asia.

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-- Managing Editor

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- Stepanov, A.I.A. - "Field practice in physical geography" by A.M. Alpat'ev, A.M. Archangel'skii, T.N. Gordeeva. p.94.
- Gornung, M.B. - "Man and volcanoes" [in French] by Edgar Aubert de la Rue. p.96.
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- Kazhibayeva, P. T. - Granulometric composition of rocks of the upper Paleozoic sedimentary complex in the Dzhezkazgan-Ulutau region. p.32.
- Kunaev, D. S. - Geology of ore bodies in the Pokro-Yugo-Zapad sector of the Dzhezkazgan deposit. p.41.
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- Ionich, V. K., Nurlybaev, A. N. - Academician Evgraf Stepanovich Fedorov; on the 40th anniversary of his death. p.98.
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